NORSK GEOLOGISK TIDSSKRIFT

UTGITT AV
NORSK GEOLOGISK FORENING
OSLO

MED STATSBIDRAG OG MED BIDRAG
FRA NORGES ALMENVITENSKAPELIGE FORSKNINGSRÅD

REDAKTØR:
NIELS-HENR. KOLDERUP

BIND 41

A/S John Griegs Boktrykkeri
Bergen 1961

INNHOLD

Contents

Side
Page

Avhandlinger

Papers

CHRISTIE, OLAV H. J.: On sub solidus relations of silicates. I. The lower breakdown temperature of the åkermanite gehlenite mixed crystal series at moderate water pressure 225
DITRICH, R. V.: Contributions to the mineralogy of Norway. No. 12. Hisingerite in "dark" larvikite 95
GEIS, HANS-PETER: Strukturelle iakttagelser ved noen norske kisforekomster. *Structural observations on some Norwegian sulphide deposits (English summary)* 173
HEIER, K. S.: Layered gabbro, hornblendite, carbonatite, and nepheline syenite on Stjernøy, North Norway 109
MICHOT, JEAN: The anorthositic complex of Håland — Helleren 157
MORTON, R. D.: Synthetic chlorapatites and equilibria in the system calcium orthophosphate — calcium chloride 223
— Contributions to the mineralogy of Norway. No. 9. On the occurrence of two rare phosphates in the Ødegården apatite mines, Bamble, South Norway 233
NEUMANN, HENRICH: The scandium content of some Norwegian minerals and the formation of thortveitite, a reconnaissance survey 197
— (se SÆBØ) 247
NIKOLAISEN, FRANK: The Middle Ordovician of the Oslo region. 7. Trilobites of the suborder Cheirurina 279
OFTEDAL, IVAR: Remarks on the variable contents of Ba and Sr in microcline from a single pegmatite body 271
RAMBERG, HANS: A study of veins in Caledonian rocks around Trondheim Fjord, Norway 1
SMITHSON, S. B.: A regional gravity study over the Permian Bærum cauldron of the Oslo region 211
SPJELDNÆS, NILS: Ordovician climatic zones 45
SÆBØ, PER CHR.: Contributions to the mineralogy of Norway. No. 11. On lanthanite in Norway 311
— and NEUMANN, HENRICH: Contributions to the mineralogy of Norway. No. 10. On synchisite in Norway 247

Notiser

Notes

Christie, Olav H. J.: On sub solidus relations of silicates. II. Preliminary note on the chemistry of melilites 322
Hessland, Ivar: International Association of Sedimentology 324
Oftedahl, Chr.: Vulkanologisk symposium i Italia september 1961 93
— Den 12. kongress i den Internasjonale Union for Geodesi og Geofysikk (IUGG) 93
Rosenqvist, Ivan Th.: Refugieproblemet og de kvartærgeologiske metodene 319

Ny litteratur

Reviews

Oftedal, Ivar: *Jac. Berzelius Brev* 325
Barth, Tom F. W.: *D. D. Bloss:* An introduction to the methods of optical crystallography 325
— *Walter Fischer:* Gesteins- und Lagerstättenbildung im Wandel der wissenschaftlichen Anschauung 326

Norsk Geologisk Forening

Møtereferater 85, 327
Årsmelding 1960, regnskap 87
Bergens Geologiske Klubb, årsmelding 1960 90
Trondheims Geologiske Klubb, årsmelding 1960 91
Medlemsliste 331

Geolognytt 328

A STUDY OF VEINS IN CALEDONIAN ROCKS AROUND TRONDHEIM FJORD, NORWAY

By

HANS RAMBERG
(University of Chicago)

Abstract. The mode of occurrence and mineral paragenesis of quartz — calcite veins in Caledonian metamorphics around Trondheim fjord are described and discussed. Quartz and calcite are the major minerals; quartz occurring in all veins, calcite in the majority of them. Chlorite, albite, biotite, muscovite, tourmaline, epidote and a few ore minerals are present in some veins in decreasing abundance.

The localization of the veins is in part controlled by the mechanical behavior of the country rocks in relation to the stresses which have prevailed during the regional metamorphism of the area.

It is concluded that the veins have been emplaced at a very slow rate contemperaneous with the regional recrystallization in the area, partly as tension fracture fillings, partly as volume-by-volume replacement bodies.

The vein minerals appear to have grown in pace with the dilation of the fractures so that empty or fluid-filled openings were hardly existing at any stage during the evolution of the veins.

Some of the elements of the vein minerals are extracted from the immediate adjacent rocks; the source of other elements is less certain.

The typical vein paragenesis in greenstone host rocks is quartz, calcite, chlorite and albite. It is shown by thermodynamics that such a mineral assemblage would tend to develop at sites of tensile — or minimum compressive — stress provided CO_2 is a very mobile constituent and the mineral assemblage in the host rock in bulk is actinolite, albitic plagioclase and quartz.

General.

The rocks in which the studied veins occur are greenschists, various types of slate, calcareous sandstone or graywacke, conglomerates and limestone. The greenschists are chiefly altered lavas and tuffs; relic pillow structure is rather common. The slate is in places carboniferous,

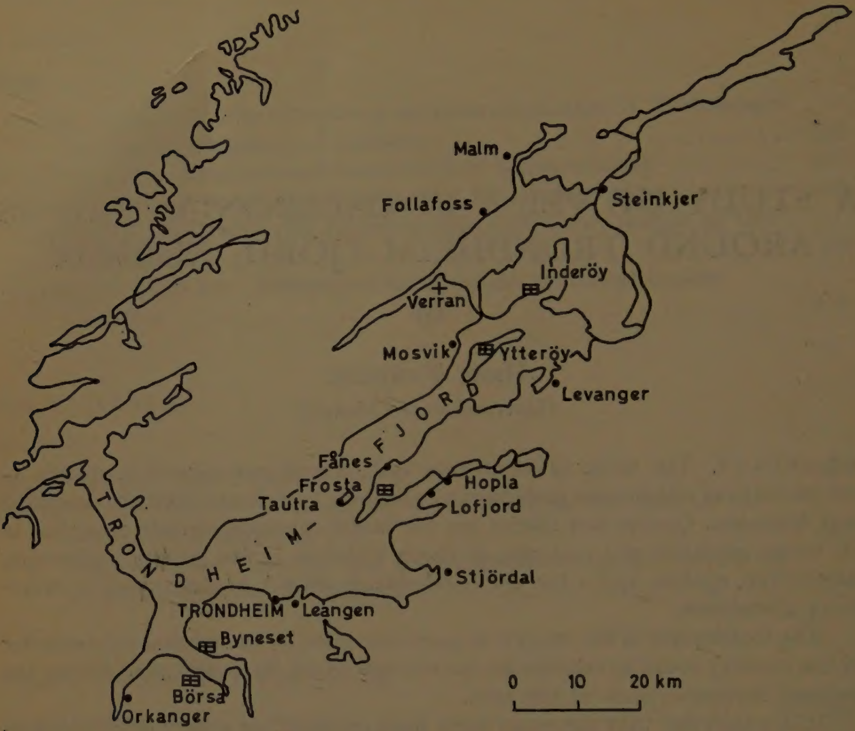


Fig. 1. Map over the Trondheimfjord area.

rich in pyrite and may contain carbonate concretions. All rocks in which the veins occur are deformed and metamorphosed during the Caledonian orogeny.

The grade of metamorphism is low in the studied region, everywhere below amphibolite facies, as indicated by lack of diopside in the greenschists (except as rare relics), but presence of chlorite and actinolitic hornblende. Garnet is generally lacking in the argillaceous metasediments whereas biotite sometimes has developed.

Quartz-calcite veins occur in all above-mentioned rock types although the abundance of veins varies from rock to rock. This is evident in complexes of intercalated slate and sandstone, for example, where veins are usually concentrated in the sandstone beds. Along most of the well exposed shore zone the veins are rather evenly scattered except for the effect due to the preference of veins to certain types of host rock. However in the W-SW part of the region, along the shore of Bymarka penninsula, veins are practically non existent in the huge masses of greenstone exposed there.

Farther South-West, in the Viggen — Orkdal area, quartz-feldspar pegmatites and aplites rather than quartz-calcite veins are numerous in the garnetiferous micaschist which constitutes the bed rock here.

Size, shape and internal structure of veins.

Most veins found in the area are small, rarely more than fifty cms. wide and some twenty meters long. More commonly the veins are less than twenty cms. wide and less than ten meters long. The shape may be planar with parallel borders, lenticular, pinching and swelling, ptygmatically folded, or irregularly branching.

Many quartz-calcite veins are zoned, usually with calcite along the borders and quarts in the center; the opposite is rare. In zoned albite containing veins, albite occurs along the border and calcite and/or quartz in the center. A haphazard distribution of minerals also occurs in a number of veins. Open vugs are very rare.

Some veins exhibit a banded structure parallel to the borders. One type of banding is brought about by stringers and elongate relics of host rock, a second type is caused by zones of shearing and granulation parallel to the walls, and a third type seems to be fillings of late minerals in fractures parallel to the borders. Similar features are well known from quartz veins in other regions (BATEMAN, 1951).

The type of banding caused by remnants of host rock is most abundant. It is usually found in quartz veins in mica-chlorite schists (slate) and consists often of thin zones rich in muscovite as «coating» between quartz grains or as lined-up inclusions in quartz. Here and there along these micaceous zones small inclusions of host rock occur. These zones represent incompletely replaced host rocks, see Pl. 1, Fig. F. It is interesting to note that small stylolite seams now and then are found in these bands of partly replaced host rocks, see Pl. 1, Fig. E.

An impression of banding is sometimes brought about by zones of granulation and mylonitization, which give rise to alternating fine grained and coarse grained layers along the veins.

In a zoned albite-calcite vein (No. 36520) in greenstone a sort of banding is caused by stringers of late quartz running through the

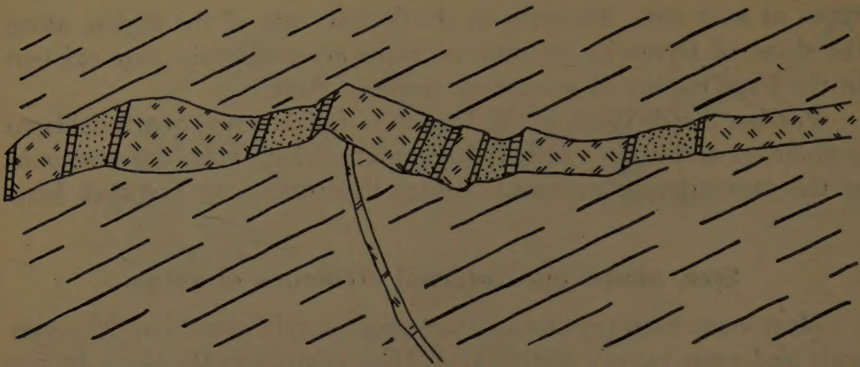


Fig. 2. Ladder vein of calcite with quartz fillings, in slate. Cross-hatched: calcite vein. Dotted: quartz. Short crosslines adjacent to quartz: recrystallized calcite. Long lines indicate schistosity in slate. Thickness of ladder vein varies from 1 to 2 cm. Lofjord.

border zone of albite parallel to the wall. At places the quartz band cuts through broken single crystals of albite, indicating that the quartz band represents a late filling of a fracture parallel to the vein.

Minute secondary veinlets of various kinds are abundant in some of the studied veins. These veinlets are generally less than one milimeter wide, cut the main veins at obtuse angle, and do not usually extend outside the main vein. Some of the secondary veinlets are nothing but zones of clear quartz through grains of dusty quartz, the dust being minute inclusions of various kinds, Pl. 1, Figs. A and B. The clear quartz of the veinlets is in optical continuity with the adjacent dusty quartz such that where a veinlet cuts across a quartz grain the clear quarts and the dusty quartz on either side constitute a single grain. Here and there small fractures occur in the clear veinlets, indicating perhaps that the veinlets developed along fractures which now only occasionally remain unhealed. Calcite is present in some of the veinlets of clear quarts. The clear quartz then exhibits crystal faces toward the calcite which occupies the central portion of the veinlets. Euhedral inclusions of clear quartz also occur within the calcite, Pl. 1, Fig. C. It seems likely that the clear quartz simply is overgrowth on the dusty quartz and probably formed in dilating cracks. Where calcite formed simultaneously with the quartz the latter developed crystal faces toward the calcite; but if calcite did

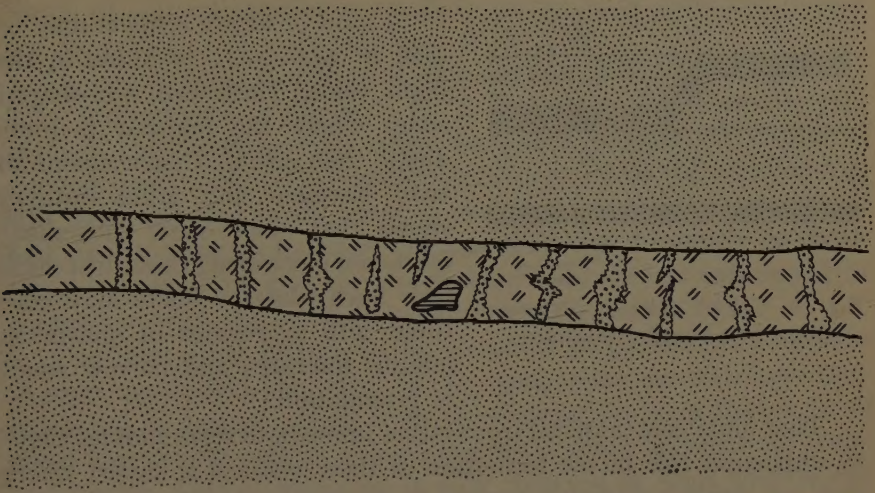


Fig. 3. Ladder vein of calcite with quartz fillings in calcareous sandstone. Thickness of vein about 1.5 cm. Lofjord.

not form the clean overgrowths of quarts along opposite walls of the fissure would merely weld together in optical continuity.

Quartz veinlets often penetrate calcite grains in a zig-zag fashion along cleavage planes in the calcite grains.

The quartz veinlets which cut across certain tape-worm calcite veins described below (p. 20) are interesting. The grey tape-worm calcite veins, which mostly occur in shale, are cut by sets of rather evenly spaced quartz veinlets as shown in Figs. 2 and 3. Along the quartz veinlets the calcite has recrystallized to a coarser aggregate. An interesting feature is that some single quartz grains extend across the entire veinlets rather than terminate in its middle, Pl. 2, Fig. E. Various circumstances (to be discussed later, p. 25), show that these cross cutting quartz veinlets grew in dilating tension cracks caused by differential plastic flow of the adjacent schist along the competent old calcite veins.

Small veinlets of calcite (without margins of clear quartz) are abundant in some quartz-albite veins. Some of the calcite veinlets are volume by volume replacement features as indicated by the fact that they selectively replace twin lamellae of albite grains as shown in Fig. 4.

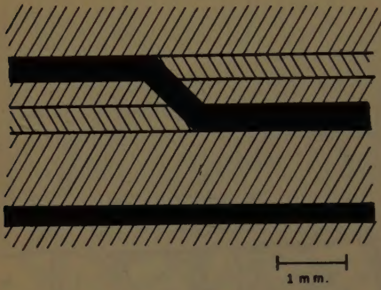


Fig. 4. Calcite (black) replacing albite twin lamellae in albite.

A kind of «dust» veinlets may be seen in aggregates of albite in the main veins. These are vein-shaped zones where the feldspar is more dusty than usual. Displacement of twin lamellae and grain outlines shows that some shear movement has taken place along these dust veinlets. Such «dust» veinlets may continue into aggregates of quartz as minute fractures and narrow zones of granulation.

That mobilized material from rather immediate surroundings plays an important role in the formation of these veinlets is shown where they cut across host-rock inclusions in the main veins. In one locality, for example, the secondary veinlets consist of a zeolite, quartz and some recrystallized albite where they cut across feldspar aggregate, but immediately upon penetrating into inclusions of greenschist, chlorite developes as a main mineral of the veinlet, Pl. 1, Fig. D. Incidentally, chlorite is a rather common mineral in some secondary veinlets within the main veins. Chlorite sometimes fills tension fractures in albite veins which have been stretched in deformed greenschist.

Albite is also a major mineral in some secondary veinlets. In veins of quartz, red albite and dark green chlorite, which occur in graywacke or sandstone between Fånes and Falstad, small fracture fillings mainly consisting of red albite are numerous.

The miniature features here described seem remarkably similar to the larger so-called fracture fillings in big pegmatite bodies as described for example by Cameron et. al. 1949. It is concluded by the present author that the miniature fracture fillings and secondary veinlets as described in this paper, are localized in the main veins because of a difference in mechanical properties of the main veins and their host rocks. The coarse grained quartz-calcite (feldspar) veins are competent and brittle relative to most of the adjacent rocks which are greenschists, black shale, slates, graywackes, etc.; consequently fracture developed in the veins while the host rocks yielded in a coherent plastic manner. Such fractures function as sites of consolidation of mineral substances which happened to be mobilized during the dynamothermal metamorphism of the area.

In several veins remnants of replaced host rock occur. A particularly striking case is shown in Pl. 2, Figs. C and D, in which a fine grained chlorite-muscovite schist has been replaced to a coarse grained quartz vein. The small-folded structure of the old schist is still well preserved in the form of undulating stringers of muscovite and chlorite through the large quartz grains. Other example on structures indicative of host-rock replacement is already mentioned in connection with the banded structure of some veins.

One interesting feature is the bleaching of greenschist inclusions in some quartz-feldspar veins. The ferromagnesian minerals have been extracted to various extent from these inclusions such that some of them now consist only of a small grained suggery aggregate of albite and quartz. This phenomenon is only observed in veins with biotite and/or black tourmaline; the extracted iron and magnesium have apparently been utilized in these minerals. The same kind of bleaching can also be seen in greenschist host adjacent to veins rich in biotite or tourmaline, see Table 1. Incidentally, strong bleaching is often seen around tourmaline veins in other areas also (RAMBERG, 1956).

Table 1. Spectroscopic analyses of bleached vein contacts in greenstones.

	19565 Unaltered host rock	19565 Bleached contact	19566 Unaltered host rock	19566 Bleached contact	19566 Intermediate between unalt. host and contact
SiO_2	46	54	38	61	55
TiO_2	2.0	2,2	3.6	1.05	1.9
Al_2O_3	16	15	18	20	18
Fe	10.5	4.9	15	1.9	4.3
CaO	6	5.3	4.1	2.2	2.6
MgO	5.1	1.7	7.9	0.48	1.35
V_2O_3	0.063	0.038	0.068	0.011	—
WO_3	0.033	0.035	0.051	0.034	0.03
MnO	0.15	0.10	0.26	0.05	0.10
Na_2O	4.8	7.4	2.7	9.6	8.4
K_2O	0.5	0.35	0.5	0.4	0.75
H_2O+CO_2	3.0	2.5	6.3	—	1.8

Analyst: O. Joensuu, 1955.

Some host-rock inclusions are associated with clusters of small grained chlorite which not only seems to grow on the rock inclusions but also occurs as well developed euhedral grains enclosed in surrounding quartz, feldspar or calcite. At places the enclosed chlorite crystals, which usually are wormshaped and elongate perpendicular to basis (see Pl. 2, Fig. A), are concentrated in quartz or feldspar to the extent that these minerals become dark green in color.

The inclusions of euhedral chlorite in quartz, feldspar and calcite offer an interesting problem: did the chlorite replace the host minerals or *vice versa*, did host and inclusions grow simultaneously, or are there some other way of explaining the intergrowth? It seems unlikely that quartz (or feldspar, or calcite) replaced chlorite because the chlorite inclusions are often perfectly developed euhedral crystals. It is perhaps not impossible that chlorite inclusions replaced the host mineral; but a more likely explanation is that chlorite inclusions and host mineral grew simultaneously, and that chlorite has greater tendency to develop crystal faces than either quartz, feldspar of calcite. The fact that chlorite occurs in the form of numerous small crystals in the host rather than as one or a few large grains or clusters may be explained by relative rate of growth of chlorite and host mineral. If the chlorite substance was supplied less rapidly than for example quartz to the site of growth, the fast growing quartz would circumspand and enclose any slow growing chlorite grain which nucleated on the quartz surface. A continued supply of chlorite substance would, however, cause nucleation of new chlorite crystals which in turn would be enclosed by fast growing quartz. The same picture may be applied to the inclusions of euhedral chlorite in feldspar and calcite.

In some cases chlorite substance may become conveyed more rapidly to the scene than quartz (or feldspar and calcite) and monomineralic clusters of rather small grained chlorite form.

Composition of veins.

Quartz and calcite are by far the most common constituents of the veins. In some veins calcite is lacking and in other quartz, but in most veins both these minerals occur together. Chlorite is probably third in abundance. Many veins in greenschist and some in graywacke are rather rich in albite, a mineral not found in veins in slate or limestone. Biotite is abundant in some veins in greenschist but is not seen in veins in other rock types. Epidote is rather rare: it occurs in some veins in epidote carrying greenschists and in small veinlets in epidote rich concretions in meta graywacke. Actinolite is a rare mineral in veins in greenschist. Sericite occurs as remnants of altered «schlieren» of host rocks in some veins in slate. Rutil is encountered in one locality. Pyrite, pyrrhotite and chalcopyrite are now and then found in veins in greenschist, so is magnetite. Black tourmaline occurs in a few veins in greenschist.

Quartz

The vein quartz is milky or dirty gray in color, mostly due to clouds of minute fluid inclusions. In places numerous inclusions of chlorite or biotite make the quartz dark green or black.

Under the microscope the vein quartz almost invariably exhibits undulatory extinction and contains numerous fluid inclusions which mostly are lined up in rather straight planes. With very few exceptions crystal faces are not developed.

The undulatory extinction is of the usual type of minute angular displacements of smaller or larger domains elongate approximately parallel to the *c*-axis. More severely strained quartz grains, however, are altered to an irregular mosaic of roughly equidmensional domains between which minute opaque inclusions often are concentrated. In some instances a thin irregularly branching coating of opaque material occurs between the domains. In quartz with relatively large inclusions of chlorite and biotite undulatory extinction is amplified around the inclusions. This may perhaps be due to stress developed around the inclusions during growth, or concentration of stress during tectonic deformation because of unlike mechanical properties of inclusions and host quartz. (See also RAMBERG, 1961).

The fluid inclusions are chiefly rounded or irregular; negative crystalls are rather rare. Many inclusions contain gas bubles, others appear to be homogeneous.

The larger fluid inclusions are generally arranged along planes which are sub parallel and extend through several grains without displacement. Several sets of such inclusion planes may be found in the same quartz grain or in the same group of grains. The planes are not related to any prominent crystallographic direction.

Some quartz grains are evenly clouded by numerous minute inclusions some of which, at least, are fluid. Because of their small size, it is impossible to identify most of these cloudy inclusions.

Small vein-shaped zones of clear quartz run through such cloudy quartz grains. The clear quartz is in perfect optical continuity with the cloudy quartz on either side of these phantome veins (see p. 4) which run across several grains without offsetting of either the veins or the grain interfaces. The veins are not related to crystallographic directions in the quartz.

In some quartz grains alternating lamellae of clear and brown cloudy quartz are seen, Pl. 2, Fig. B. The brown cloudy lamellae, some of which under high power show up as zones rich in minute fluid inclusions, have lower index ot refraction than the clear quartz. The border between the two kinds of lamellae is sharp, the lamellae being in optical continuity without angular displacement. It is probable that the low index of the brown lamellae is a mass effect of minute submicroscopic fluid inclusions. In parts of the lamellae where the fluid inclusions are large there is no difference in refractive index of clear and brown lamellae. A similar explanation has been offered by Folk and Weaver (1952) for the low index of refraction of chalcedony.

Rows of larger inclusions often run across the brown lamellae. Such rows of larger inclusions do not occur in the intercalated clear quartz, yet they can be traced without offsetting through several brown lamellae. This indicates that shearing parallel to the lamellae was not very pronounced during the origin of this kind of structure.

Most of the lamellae make a small angel with the basal plane, only in very few cases has an angel of more than 45° between lamellae and basal plane been observed. This agrees with the attitude of similar lamellae described by Ingerson and Tuttle, 1945. In contrast to what is observed here, however, Ingerson and Tuttle state that the lamellae have higher index of refraction than the clear quartz. Fairbairn, 1941, describes lamellae with lower index than the clear quartz.

The brown color of the lamellae under the microscope is probably due to the scattering of short-wave light by the liquid inclusions such that mostly light of yellow and red color passes through the lamellae. (C. R. Pelto 1956).

Quartz appears to have grown during most of the history of formation of the veins. It is particularily obvious that it usually continued to grow later than most other minerals through which it forms fracture filling and replacement veinlets. Occasionally chlorite and sometime calcite have crystallized later than even late quartz.

Albite

The only feldspar occurring in some of the veins is albite. Albite is restricted to veins in greenschists (meta basalts, often with relic pillow structure) and much more rarely to some veins in graywackes or impure sandstones. The feldspar in the veins in the greenschists is always white or green (due to inclusions of chlorite) and often occurs as well developed euhedral crystals particularily when floating in «clusters» of chlorite in the veins. The albite which occurs in veins in the meta graywackes or sandstones is always reddish.

Albite usually constitutes the border zone in zoned veins in greenschist thus indicating an early stage of formation. However, albite may also be found as well developed euhedral crystals (themselves often containing euhedral chlorite inclusions) in clusters of chlorite which is a rather late mineral in most veins. On the other hand fractured albite grains are now and then healed by chlorite. One can also find secondary veinlets of albite which is clearly younger than chlorite in some veins. There seems to be no general sequence of crystallization of the various vein minerals.

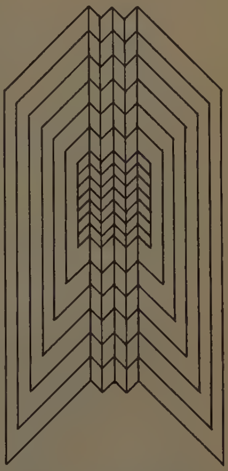


Fig. 5. Albite twin lamellae along middle of albite crystal. Densly lined core in the nucleus formed at relatively high supersaturation, the concentric zones indicate layers added by growth at small supersaturation.

Most albite crystals have a few narrow lamellae of albite twins along the middle, Fig. 5. It seems that the feldspar started to grow as a twinned nucleus but continued to grow in untwinned state. This could be due to rapid growth for a relatively short time just after the nucleation, and a subsequent slower growth. Nucleation of a crystalline phase requires a certain degree of supersaturation. At the moment the nucleus forms the environment has therefore a high concentration of active albite constituents and a «rush» growth occurs. During this

Table 2. Spectroscopic analyses of red and white albites.

	SiO_2	Al_2O_3	Fe	MgO	CaO	K_2O	Na_2O
Red Ab, 36581 ...	70	19.5	0.035	0.04	1.2	0.1	11
White Ab, 36533 .	71	18.5	0.06	0.2	1.5	0.2	11
Red Ab, 36584 ...	72	18	0.06	0.05	2.2	0.3	10.2

Analyst: O. Joensuu, 1955.

«rush» growth twins, — which represent a little higher energy than untwinned crystals — would probably form for similar reasons that for example high-energy metastable disordered crystals tend to grow in supersaturated environments. After a while the «rush» growth causes sufficient decrease of concentration to make subsequent growth slow enough to develop stable untwinned feldspar. This is only true for growth in directions perpendicular to the interface of the albite twins; the originally formed twinned lamellae continue to expand parallel to the intergrowth surfaces as shown in Fig. 5.

According to analyses there is no distinct compositional difference between the white and the red feldspar, both are rather pure albite (Table 2). Under the microscope, however, the red feldspar is dusty due to minute opaque inclusions. It is interesting that the two types of albite are so completely dependent upon the type of host rock.

Chlorite

occurs in most of the studied veins. The mineral is mostly found in the form of rather pure monomineralic clusters in the veins or as euhedral worm-like inclusions in quartz, feldspar and calcite. Chlorite may also occur in secondary fracture fillings in the main veins. The chlorite may vary somewhat from vein to vein in optical characters and color. However, within a single vein and in a swarm of veins in a given host rock the chlorite remains virtually constant at least as far as color and optical properties go.

The color of the chlorite varies from dark green to light grayish green, the index of refraction of the darkest chlorite being higher than that of the light chlorite. The optical properties of several chlorites from various veins are given in Table 3. Chemical composition as determined by spectroscopic methods are shown in Table 4.

Table 3. Optical characteristics of chlorites.

Greenschists				Slate, schist and graywacke		
Sample nos.	γ of vein chlorite	Host rock color	γ of host rock chl.	Sample nos.	γ of vein chlorite	Remarks on host rock
19555	1.628	dark green	1.628	35736	1.620	sericite schist
19557	1.630	—»—	1.630	35739	1.620	—»—
19558	1.628	—»—	—	35751	1.622	—»—
35759	1.630	—»—	1.628	35776	1.618	sericite, chlorite schist
35762	1.630	—»—	1.628	35777	1.616	—»—
35763	1.630	—»—	—	35778	1.618	sericite schist
35766	1.628	—»—	—	35788	1.618	—»—
35791	1.634	—»—	—	35789	1.618	—»—
35797	1.630	—»—	—	36531	1.620	impure sandstone
35800	1.634	—»—	1.632	36532	1.620	—»—
36520	1.624	—»—	1.628	36546	1.628	graywacke
36527	1.624	—»—	1.624	36548	1.628	—»—
36523	1.618	light green	1.620	36553	1.620	impure sandstone
36533	1.620	—»—	1.622	36584	1.620	schist
36560	1.620	—»—	1.618	36402	1.618	—»—
				36416	1.616	—»—
				36421	1.618	—»—

Table 4. Spectroscopic analyses of chlorites and biotites.

	Chlorites		Biotites		
	35777	35791	19565	19566	35767
SiO_2 %	27	25	38	37	37
TiO_2 %	0.06	0.038	1.95	1.90	1.99
Fe %	19	24.5	16.9	15.6	16.9
MgO %	18.5	11.5	11.2	11.4	10.2
Al_2O_3 %	23	21	16.3	15.8	17.6
MnO %	0.11	0.65	0.14	0.13	0.15
CaO %	—	0.03	0.46	0.23	0.16
V_2O_3 %	—	—	0.102	0.103	0.104
Cr_2O_3	—	—	—	—	—

Analyst: Dr. R. Kretz, 1958.

Empiric formulae based upon structure formulae: $(Mg, Fe, Al)_9 (Si, Al)_6 O_{21}.6H_2O$ for chlorite, and $K_2O (Mg, Fe, Ti, Al)_6 (Si, Al)_8O_{21}.2H_2O$ for biotite:

Chlorite
35777: $(Mg_{4.07}\ Fe_{3.02}\ Mn_{0.01}\ Al_{1.99})\ (Si_{3.99}\ Al_{2.01})\ O_{21}.xH_2O;\ \gamma = 1.616$
Chlorite
35791: $(Mg_{2.74}\ Fe_{4.21}\ Mn_{0.09}\ Al_{1.96})\ (Si_{4.00}\ Al_{2.00})\ O_{21}.xH_2O;\ \gamma = 1.634$
Biotite
19565: $K_xO_{1/_2x}\ (Ti_{0.22}\ Mg_{2.46}\ Mn_{0.02}\ Fe_{2.6}\ Al_{0.4})\ (Si_{5.59}\ Al_{2.41})\ O_{21}\cdot xH_2O;\ \gamma = 1.639$
Biotite
19566: $K_xO_{1/_2x}\ Ti_{0.22}\ Mg_{2.6}\ Mn_{0.02}\ Fe_{2.56}\ Al_{0.51})\ (Si_{5.66}\ Al_{2.34})\ O_{21}\cdot xH_2O;\ \gamma = 1.634$
Biotite
35767: $K_xO_{1/_2x}\ (Ti_{0.23}\ Mg_{2.28}\ Mn_{0.02}\ Fe_{2.72}\ Al_{0.67})\ (Si_{5.56}\ Al_{2.44})\ O_{21}.xH_2O;\ \gamma = 1.638$
Index of refraction determined by Mr. J. McLelland who also calculated the formulae.

The composition of chlorite in the veins is closely related to type of host rock. It is clear from field observations that chlorite is much more abundant in veins in greenschists than in veins in slates and graywacke or sandstone. It is also apparent that most veins in greenschists carry a darker chlorite than the veins in other kinds of host rocks. Table 3 shows that the index of refraction of the vein chlorite in slates and graywacke is lower than the index of vein chlorite in greenschists. There are some exceptions such as Nos. 36523, 36533, 36560, Table 3, which are vein chlorites from greenschists yet they have low index of refraction and light color. However, these veins occur in exceptionally light colored greenschists which themselves contain chlorite with low index of refraction. A few optical checks were made on the chlorite in the darker greenschists of the type in which most chlorite bearing veins occur. It is seen that this host-rock chlorite has similar high refractive index as the corresponding vein chlorite. Two vein chlorites in impure sandstone or graywacke show unusually high index of refraction. (Nos. 36546, 36548, Table 3). Unfortunately the host rocks contained too little chlorite to allow index of refraction to be measured; it is interesting to note, though, that the one of these host rocks which was analysed has very high Fe/Mg ratio; indeed, higher than any of the analysed metasedimentary rocks, (Table 5, No. 36547). Table 5 shows also that the two greenschists which have been analysed have higher Fe/Mg ratio than the metasediments with the exception of the abovementioned graywacke No. 36547. This harmonizes with the fact that vein chlorites in greenschists have higher index of refraction and higher Fe/Mg — ratio than vein chlorites in metasediments.

Table 5. Spectroscopic analyses of some host rocks.

	Greenst. 19565	Greenst. 19566	Calc. sandst. 35740	Slate 35741	Sericite sch. 35778	Sericite sch. 35779	Sandst. 35784	Sandst. 36547	Sandst. 36583	Sandst. Fa, 10,4,54	Slate Ny, 25,4,54	Grayw. La, 25,4,54	Grayw. 36552
SiO_2	46	38	45	63	42	56	71	70	73	71	55	63	65
Al_2O_3	16	18	8.2	20	21	17	14.5	14.5	14	14	17	15	14
TiO_2	2.9	3.6	0.63	0.5	1.45	0.65	0.27	0.63	0.3	0.65	1	1.6	1
Fe	10.5	15	2.4	4.5	8.6	5.5	1.5	3.1	1.35	3.3	6.3	4.4	3.3
CaO	6	4.1	17.5	2.5	0.95	2.4	1.4	3.0	2	3.3	3.8	3.8	4.4
MgO	5.1	7.9	2.1	5	9	5	0.9	1.7	1.0	2.2	5.2	3.5	2.8
V_2O_3	0.063	0.068	0.015	0.02	0.045	0.025	0.009	0.02	0.006	0.013	—	—	0.019
MnO	0.15	0.26	0.2	0.045	0.05	0.05	0.035	—	—	—	—	—	—
Cr_2O_3	—	—	0.015	0.005	0.023	0.075	0.015	0.006	0.006	0.01	0.02	0.013	—
Na_2O	4.8	2.7	2.75	3.5	2.2	3.1	3.9	3.5	4.6	3	1.9	2.4	2.6
K_2O	0.5	0.5	0.3	1.9	2.3	1.4	2.0	1.7	1	2.1	2.9	2.8	2.95
H_2O	3	6.3	18	—	6.5	5	4	—	—	—	2	4	—
CO_2													

Analyst: O. Joensuu, 1955.

Remarks: Veins in 35784, 36547, 36552, 36583 and Fa, 10,4,54 carry albite. Veins in no. 36547 contain a dark high-index chlorite, see table 3.

Biotite
occurs in veins in greenschists. Segregations of almost monomineralic biotite clusters are also found in these rocks. A bleached zone depleted on ferromagnesian minerals may surround biotite clusters or veins rich in biotite. Table 1 shows the chemical alteration in such bleached contact zones. Biotite is not found in veins in slate and meta graywacke. Analyses of vein biotites are given in Table 4.

Muscovite
occurs associated with host-rock inclusions in some veins in slate.

Black tourmaline
is present in some veins in greenschists. Such veins are often surrunded by bleached host rock, and may contain inclusions from which all traces of ferromagnesian minerals have disappeared. Tourmaline crystals are often broken, with angular displacement of the fragments and quartz filling between them.

Epidote
is found in some veins in greenschists containing themselves epidote of same character as the veins. Epidote is also found in quartz fillings in tension fractures in strongly deformed calcareous inclusions in some impure sandstones (see p. 21). In these dilation veins needleshaped single crystal of epidote often extend from border to border across the entire vein. The needles may be bent or broken.

Actinolitic hornblende
occurs in small amounts in a few veins in actinolitic greenschists.

Calcite

The only carbonate found in the veins is calcite which of course is very abundant. It generally shows sign of severe strain such as bent twin lamellae.

Oxides and sulfides

Rutil is found in one vein at Balsvik gård, Frostatangen. Magnetite, pyrite, pyrrhotite and chalcopyrite occur in small amounts in some veins in greenschist.

Structural setting.

Structural relations show that many veins formed in dilating fissures. Other veins show evidence of replacement of host rock. However, since almost every vein studied has been strained during or after their formation clear evidence of dilation or nondilation is not generally available.

Most veins unquestionably emplaced in dilating fissures are either feather joint fillings, or fillings between boudins.

FEATHER JOINT FILLINGS

Zones of moderate shear across the schistosity are common in slates and greenshists in the area. The en echelon fractures formed along such shear zones are generally filled with veins of the kind studied. Usually the shear zones are only a few decimeters to one meter wide. The resulting en echelon fractures are therefore sites of but small quartz veins.

It is noteworthy that the angel between the individual en echelon quartz veins and the shear zone often is considerably less than 45°, Fig. 6. This contrasts the 45° or greater angle between tension cracks and shear movement as found in experiments and predicted in theory. In this connection it is of interest to recall the tests of clay made by RIEDEL (1929) who reported two sets of ruptures in bodies of clay strained in rotational shear between rigid walls. One set of ruptures which made about a 45° angle with the direction of shear movement, were tension cracks oriented perpendicular to the direction of maximum elongation of the strained sample. These tension cracks rotated and assumed an angle of 60° with shear direction during the strain experiments. Another set of fissures developed at only 17° angle with the shear motion. The last set proved to be shear fracture along which slip took place. (See also RAMBERG, 1959). It seems that many of the en echelon quartz veins in the area follow such low-angle shear fractures rather than true tension cracks.

Riedel reports that the tension fractures developed earlier than the shear fractures in his experiments. In this respect Riedel's experiment does not reproduce the conditions of the strained rocks in the studied area where generally low-angle shear fracture veins are not

Fig. 6. Quartz fillings in echelon fractures in impure sandstone. Stuenget, ca. 10 km W of Steinkjer.

predated by high-angle tension fracture veins. Of curiosity the writer repeated Riedel's experiments using putty rather than clay in the shear zone. It was then found that a set of shear ruptures making about 20° angle with the direction of movement of the adjacent rigid blocks were the only fractures which developed during the tests. The discrepancy between the tests of Riedel and those of the writer is probably related to the unlike rheomorphic properties of clay and putty.

The vein filled feather fractures as described above developed during shear movement along narrow zones across schistosity and layering in schists. In some localities (between Steinkjer and Beistadsundet) feather fractures have developed in response to shear movement parallell to the layering in the schists. Here relatively thin beds of slate are intercalated with much thicker beds of sandstone, conglomerate and graywacke. Quartz veins are strongly concentrated in the slate beds most of which contain highly irregular and turbulently folded veins. In a few localities the quartz veins in the slate beds are less deformed and occur as rather straight bodies arranged in en echelon fashion along the slate beds as shown in Fig. 7. At the locality shown in Fig. 7 schistosity or cleavage in the slate makes a 20° angle with the layering and the thin lenticular quartz veins are perpendicular to the schistosity. The attitude of the schistosity in the slate layer showes that considerable displacement has taken place parallel to the layering, the competent sandstone having slid a.ong the incompetent slate beds. This sliding of the sandstone

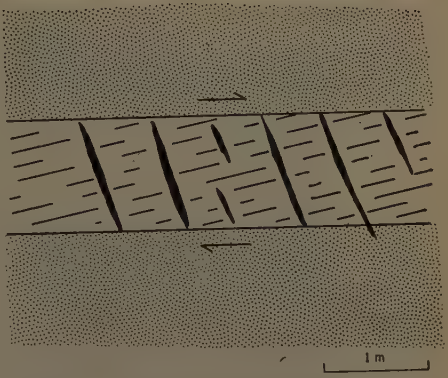


Fig. 7. Layer of slate with quartz veins sheared between two beds of sandstone. Sæterviken.

along the slate beds has caused diagonal stretching — and compression — in the slate giving rise to schistosity perpendicular to the compression and tension fractures perpendicular to the stretching. The tension cracks probably started out at about 45° angle with layering, but were slowly rotated as the strain continued. It is interesting to remember in context with the above that the attitude of the low-angle shear planes as developed in Riedel's and the writer's tests is opposite to the attitude of the schistosity in the slate layer as related to direction of movement of the adjacent sandstone. In general the acute angle between layering and schistosity in an incompetent layer situated between competent beds points in the direction in which the competent beds slide past the incompetent (Fig. 7). This has been found wherever independent signs of relative movements of beds are present such as in folds, and there are no reasons to doubt that this is true in the studied localities. In the writer's repetition of Riedel's experiments a set of linear or planar feature with the same angular attitude as the schistosity in sheared incompetent rocks were actually found. This linear feature consisted of minute folds and/or striations on the putty surface with axes less than 45° with he layering and the acute angle pointing in the direction of sliding of the wooden block past the putty layer. Deformed originally circular marks on the putty surface showed that the striation was parallell to the long axis of the strain ellipsoid. The angle between this striation or axis of microfold and the direction of sliding decreased gradually with increasing strain. The minute surface folding shows the presence of unidirectional compression in the incompetent layer. The angular attitude of the compressive strain is in accordance with the assumption that schistosity developed perpendicular to compression and/or parallell to extension rather than parallel to slip planes. See photos in Ramberg, 1959.

The quartz veins were concentrated in the incompetent slate beds because the tectonic shear movement parallel to bedding was localized in these beds thereby creating tension and shear fractures in which the veins would grow. The tectonic movement continued after some of the quartz veins were formed causing considerable folding and crumpling of the veins.

VEINS IN BOUDINAGE AND RELATED STRUCTURES

Vein filled tension gashes are numerous in layers of calcareous sandstone, graywacke, conglomerate and limestone intercalated in shale. Layers of ultrabasic rocks in greenschist (Inderøy) also contain tension joints with veins. In the region around Lofjord—Fættenfjord a striking type of ladderveins are common. Strained calcite veins are cut by numerous rather evenly spaced quartz veins. The quartz veins are confined to the old calcite veins and do not project into the adjacent host rock which usually is slate or calcareous sandstone, Fig. 2.

According to some authors ladderveins fill shrinkage cracks caused by cooling of the dike in which the ladderveins occur. This explanation is obviously not applicable to the studied ladderveins because the volume occupied by the crosscutting quartz veinlets (see Fig. 2) is in some cases more than thirty per cent of the calcite vein. No thermal shrinkage could be of that magnitude. Furthermore, in most instances the crosscutting quartz veins follow but one set of parallel tension fractures thus showing that the tension was pronouncedly anisotropic within the plane of the original calcite vein. A thermal shrinkage should give rise to isotropic tension and several sets of crossing tension joints.

It is plain that these ladderveins developed in the same manner that boudinage structure developes, *i. e.* by extension plastic flow in incompetent rocks and creation of tensile stress with resulting tension fractures in intercalated competent or brittle rocks.

In one of the collected samples the host rock to these fractured calcite veins is a pyrite-containing slate which shows clear evidence of elongation in the form of elongate — and even fractured and separated — pyrite grains and well developed pressure shadows adjacent to the pyrite crystals P. 2, Fig. F. In this sample the fractures in the calcite veins are closely perpendicular to the elongation in the host slate. It is also interesting to note that a striation parallel to

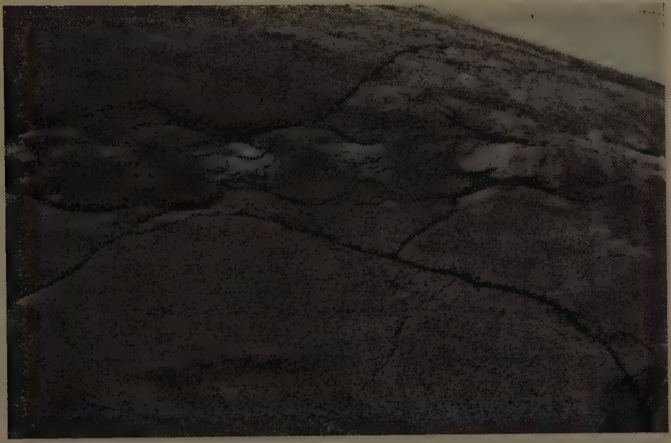


Fig. 8. Quartz-epidote fillings between boudins of deformed calcareous concretion (?) in impure sandstone. Kvistvik, ca. 15 km NE of Fånes, Frosta.

the host-rock elongation is present on the quartz fillings at their boundary toward the adjacent slate. Incidentally, the mentioned boundary surface is rather straight showing no or very little flow of host slate into the areas between the boudins of calcite vein. It seems that the interboudin cracks were continuously filled with quartz during their expansion.

In a thick layer of metamorphosed graywacke and impure sandstone at Lekshammeren between Fånes and Falstad strongly flattened epidote-rich inclusions contain numerous crosscutting quarts veinlets, Fig. 8. These veins follow one set of parallel cross joints in the flat elliptical inclusions. It is interesting that the quartz veins in these epidote-rich inclusions also contain epidote in the form of bent needles which may extend across the entire veinlets. The flattened inclusions themselves once probably were calcareous concretions. Such concretions are encountered in less metamorphosed impure sandstone between Lofjord and Fætten.

Vein filled tension joints are also numerous in boulders in conglomerat in the area.

The kind of veins mentioned above are all formed in the same general structural setting, *viz.* in tension gashes developed in compe-

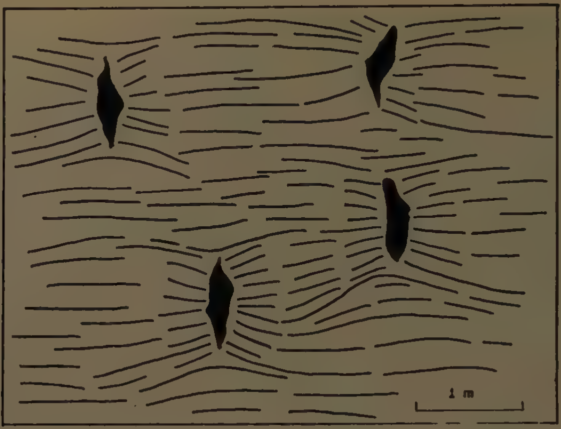


Fig. 9. Lenticular quartz veins in mica schist. Fættenfjord.

tent layers, inclusions, or old veins due to extension flowage and elongation in the surrounding incompetent rocks. In other words these veins are related to boudinage structure of one variety or other. They are accordingly formed under conditions quite different from those prevailing during the formation of the feather joint veins described above.

At several places crosscutting veins are associated with structures very similar to boudinage. In these structures there are no noticable mineralogical or textural difference between boudins and host rock. Fig. 9 shows examples of this structure. It is seen that the structure

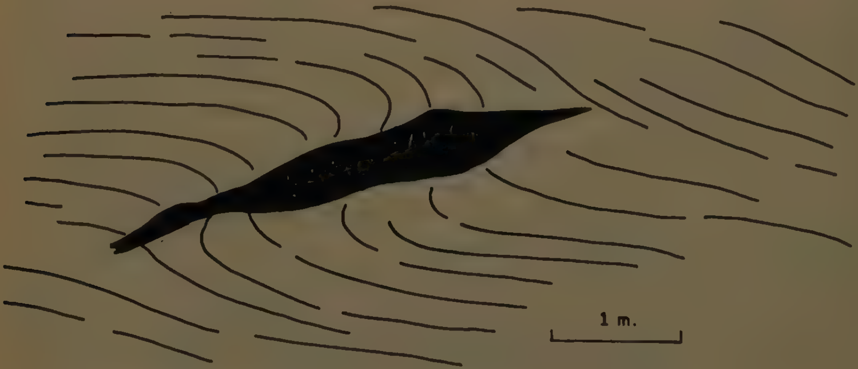


Fig. 10. Quartz vein in "drag fold". Fættenfjord.

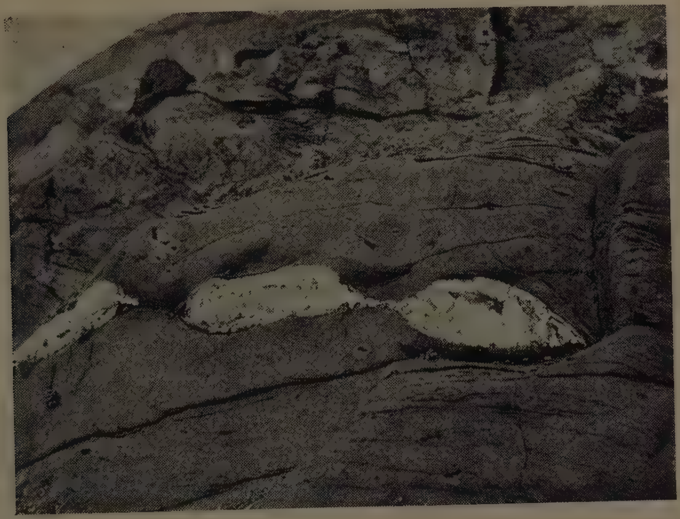


Fig. 11. Pinch-and-swell quartz vein, Frosta. Length of swells about 1 m.

of host rock close to the veins is very similar to that developed at the ends of boudins giving them their characteristic barrel shape. However, in the considered case only the one ends of «boudins» are developed as there is no distinct layer of boudins but rather scattered crosscutting lenticular quartz veins surrounded by structures similar to boudin ends.

VEINS IN DRAG FOLDS

Some veins occur in what seems to be tension gashes in drag folds as shown in Fig. 10. If the fold in Fig. 10 was formed by drag along the layering the one limb was under tension and the vein filled fissure would correspond to a tension crack.

VEINS NOT CLEARLY RELATED TO FRACTURES

Many veins occur in structural setting which seems not related to fractures. Conformable lenticular or pinching and swelling veins, Fig. 11, in schists did obviously not form in fractures if rotation and deformation are excluded. However, signs of strong deformation are indeed present in most veins and their adjacent schists. Quartz is for example always strained and often very strongly so inasmuch as bigger grains are deformed into a mosaic of smaller ones, fractures have been common as indicated by rows of liquid inclusions and phantome veins in quartz, and even stylolithes between quartz grains occur. Twinned albite grains are often bent, and calcite grains show almost invariably signs of strain. Pinch and swell structure is in the writers opinion mostly caused by compression and stretching after or during the growth of veins (RAMBERG, 1956). Ptygmatically folded veins, which are numerous in the area, are also results of compression perpendicular to schistosity and extension plastic flow in the host schists (KUENEN, 1938; RAMBERG, 1959).

All in all, therefore, it does not seem unreasonable to the writer that many of the conformable veins started to form as disçonformable bodies in tension cracks or shear fractures. Tectonic deformation during or after the growth of the veins would cause plastic flowage in incompetent schists in which the veins would behave to some extent like competent inclusions. The veins could be rotated to become conformable to the host schist, and/or a new conformable schistosity could develop locally around the competent vein bodies. During this process some recrystallization will take place in the vein such that mylonitization and other cataclastic deformation need not be pronounced. Various steps on this evolution is actually found in one part of the area. In the above-mentioned (p. 19) slate beds between thick sandstone beds in the Steinkjer—Beistadsund region, different degree of deformation is visible. In the slate beds which have been only moderately strained (see p. 00) quartz veins are straight cross-cutting fracture fillings, Fig. 7. In the more severely deformed slate beds, however, the quartz veins are irregularly folded, or pinching and swelling. These veins are commonly conformable to the schistosity of the slate.

In concluding this section it should be made clear that though many veins clearly are related to fractures of one kind or other, it is not always true that they simply represent fracture fillings. Many evidences of replacement of host rock are found such as shown in Pl. 2, Fig. C. Thus accompanying the filling of fractures (which probably never were empty but rather expanded in pace with the growth of vein minerals, see p. 31) replacement of host rock adjacent to the fracture took place in some instances.

MECHANISM OF EMPLACEMENT OF THE VEIN MATERIAL

The field observations indicate that the studied veins are results of metamorphic differentiation. In other words the veins are contemporaneous with the regional metamorphism in the area and developed as a result of rearrangement of matter during the metamorphism. The fact that the veins are confined bodies without visible avenues of transport from distant sources points toward such an origin. So does the observation that many vein minerals reflect the chemical and mineralogical composition of the host rocks. Moreover, the mineral assemblages show that the temperature at which the vein minerals formed was very closely the same as that prevailing in the host rocks during their metamorphism. (Greenschist — to low epidote amphibolite facies). Although the idea that many veins form by metamorphic differentiation is old in geologic literature, the mechanism of emplacement of such veins is open to discussion.

Interpretation of the structural pattern in the field show that a majority of the veins occur at sites where compressive stress has been at local minima (or tensile stress at maxima) during the tectonic evolution of the area. Of course, ore geologists have realized long ago that this is often true for ore veins, but the physical reason for the localization of veins in such structural sites is not generally realized.

It seems reasonable to consider two major kinds of emplacement processes of the vein minerals: 1) precipitation from bodily flowing solutions, or colloidal suspensions, a process which many ore geologists probably would favor; 2) diffusion of activated individual particles in a more or less stagnant pore solution or intergranular film, through mineral lattices and their fault network, and along surfaces and interfaces.

For reasons discussed below the conventional idea of a flowing solution or suspension as carrier of the vein minerals is not acceptable to the present writer. In dealing with such carriers one may visualize two different conditions in rocks:

1) *Underpressure solutions:* The rock is a rigid porous or fractured framework in which the moving solutions are under lower pressure than the minerals in the compact parts of the rocks. This is probably the usual situation in rocks carrying ground water and mineral oil.
2) *Equalpressure solutions:* The rock is plastic to the extent that the

openings tend to collaps (by recrystallization, crystal slip, etc.) until the fluid solution is under the same pressure as the minerals.

The underpressure solution model is thermodynamically unstable because the openings will tend to clogg by consolidation of mineral substances which have been activated or dissolved at points of high compressive stress. Provided any chemical processes take place in the rock at all, this clogging will go on until the pore solution is forced out of the system or until shrinkage of the pore volume has increased the liquid pressure to the same intensity as the rock pressure. The situation has then become identical to case 2 which will be discussed below. The clogging effect can not only be predicted theoretically (Riecke's principles), but it has also been demonstrated experimentally by FAIRBAIRN (1950) and the present author.[1] A clogging of the kind described is obviously more rapid for small openings than for large ones, and it seems that any microscopic conduit in rocks is exceedingly unreliable as channelway for flowing underpressure solutions called upon to precipitate larger bodies of vein minerals. The minute conduits must vanish long before the larger vein chambers are filled.

This is of course not true for unmetamorphosed sandstone, limestone etc., but the process goes on in any compacting sediment, it is of major importance in diagenesis, and as soon as regional metamorphism sets in the clogging process is complete thus rendering metamorphic rocks impermeable to flowing solutions. This is shown by the common occurence of «late» minerals in pressure shadows even in low grade slate, (Pl. 2, Fig. F) by the widespread healing of cracks in metamorphics, as well as by the fact that anisotropic structure developes readily in stressed rocks as a result of recrystallization. In regional metamorphic rocks only extensive macroscopic fissure systems such as faults and breccia zones are potential channelways for flowing fluids. However, the studied quartz-calcite veins are confined bodies of moderate size not connected with such fracture systems. We cannot escape the conclusions, then, that the vein material was supplied through the bulk of seemingly compact host rocks. As this could not occur in the underpressure pore solution model we shall turn to the equalpressure pore solution model.

---

[1] Recent experiments with sodium chloride at differential pressure have shown that the crystals dissolve in the high-pressure chamber in the bomb, and the dissolved ions diffuse toward the low-pressure chamber where precipitation occurs.

In the equalpressure model the fluid solution is under the same pressure as the rock minerals and therefore prevents the pores from clogging by mineral growth or collapsing by plastic flow. (Incidentally, this is the reason that fluid inclusions are stable in metamorphic minerals. The pressure in the fluid inclusions has become equalized with the pressure in the host mineral lattice — plus an effect due to the interfacial tension — and there is no further tendency for the fluid inclusions to shrink neither by plastic adjustment in the host mineral nor by growth of mineral substance along the circumference of the fluid inclusions. However, such fluid inclusions in metamorphic minerals are not interconnected and cannot therefore function as conduits for flowing fluids.)

We may now ask how the equalpressure model could account for precipitation of veins minerals at sites of low compressive stress? It surely cannot be a question of temperature difference. Why should local temperature minima in metamorphic rocks coincide with local stress minima? Precipitation due to mixing of two (or more) chemically incompatible fluids at the vein site seems too improbable to hold as a general explanation. It would be more reasonable to assume either that the solutions are somewhat supersaturated prior to entering the vein, or that the solutions become supersaturated upon entering the vein site because of the low compressive stress here.

Now, if the solutions are originally supersaturated precipitation would take place in the minute pores and fissures thus causing clogging as in the underpressure model, and rendering rocks impervious before larger fractures would be filled with vein minerals. One recalls in this connection that the solubility of a crystal depends upon the curvature of its surface in such a fashion that the solubility in a pore fluid of a porous crystalline aggregate decreases with radius of curvature of the pores. Hence small pores or micro tunnels may clogg by mineral growth at concentrations too low to cause crystallization in larger openings.

Discarding then the possibility of original supersaturation of vein forming solutions, we shall consider the drop in pressure upon entering the vein siste as a cause of mineral precipitation. Whether a saturated solution shall become supersaturated or undersaturated upon entering a potential vein site depends upon the effect of pressure on solubility. For some solids the solubility in aqueous solutions increases with increasing pressure but for other solids the opposite is true. It is not known how the various vein minerals behave in this respect, but it is improbable that the solubility for every one of these minerals should increase with rising pressure thereby making the low pressure at the vein site a plausible cause of precipitation. Such a situation is particularily unlikely in view of the fact that every one of the rock forming minerals in the area also occur in veins. Let us, however, nevertheless consider a mineral whose solubility rises with increasing pressure and analyse what happens in a potential vein site such as the openings between boudins in Fig. 9. [It is very convenient that the system quartz-water behaves in this way within a certain P,T-field.]

When the crack starts to open the pore solution flows into the crack and silica will precipitate as long as the pressure on the fluid in the crack is less than the pressure on the fluid in the rock. In terms of volume, however, the precipitated amount will only occupy a small part of the opening because the solubility of quartz in aqueous solution is very limited and the decrement in solubility with a finite drop in pressure is only a small fraction of the total solubility. Pore solution from the adjacent rocks will flow into the crack until the pressure of the solution in the vein opening becomes equal to that of the mobile pore solution in the adjacent rock. According to our earlier assumption the pore solution and the rock minerals are under the same pressure so we end up with a situation that the pressure of the fluid solution in the crack is identical to the rock pressure. There is therefore no further tendency for solutions to flow toward the crack save for the amount necessary to maintain constant pressure in a crack if it continues to expand. It is worth noting that under the equalization of pressure between fluid filled crack and bulk of adjacent rock the early formed quartz in the vein will dissolve because solubility increases with rising pressure. Thus if this was the right model of vein formation one should expect to find numerous cracks and fractures in metamorphic terrain which are essentially empty (the fluid could have been lost during late fracturing when the complex was elevated to the surface) or fluid filled except perhaps for thin crustasions along the border. Possible collaps of the opening by plastic flow would leave detectable signs. As is well known, this is not what one encounters in such rocks where cavities are exceedingly rare not only in the studied area but in regional metamorphic terrain in general. It seems to the author that any realistic model of the vein forming mechanism must account for the established fact that fractures and other potential openings in metamorphic rocks are filled with minerals rather than fluids. As has been pointed out in several papers dealing with pegmatite veins. (RAMBERG, 1945, 52, 56), diffusional transport of mineral matter satisfies this requirement because in this model any local low-pressure region functions as a sink for the diffusion current not only attracting mobile constituents in the surrounding rock complexes, but also causing the activated constituents to unite to mineral species. (See also REITAN, 1958, 1959).

There are two physical theorems which are especially important for the understanding of the diffusion model of vein formation: One theorem is that at constant temperature, diffusion is driven by free-energy gradients: diffusion of a given constituent always goes from high to low partial free energy of that constituent. (Incidentally, this is not necessarily true when temperature is nonuniform.) A concentration gradient may be regarded as the driving force of diffusion only when it is parallel to the corresponding free-energy gradient. Wherever a concentration and a free-energy gradient of a given constituent point in diverging directions, such as could easily happen in rocks under anisotropic stress, it is always the free-energy gradient that directs the diffusion current.

The second theorem is that the total free energy of a crystal always increases with increasing pressure. This is also true for the partical free energies of each constituent of a crystalline compound provided that they have positive partial volume. It is unlikely that any of the major constituents of minerals have negative partial volume for this would mean that the volume of the mineral would shrink if the constituent in question is added to the mineral. Such a situation could be imagined for a small atom (ion) which occupies empty sites in an open mineral structure. It is for example perhaps conceivable, but hardly probable, that sodium has a negative partial volume in hornblende where sodium occurs in the otherwise empty X-position: $X\,Ca_2(Mg,F)_5(Si, Al)_8O_{22}(OH)_2$. Although some elements may have negative molal volume in certain minerals it is always true that the total molal volume of every mineral is positive; consequently increasing pressure increases the sum of the partial free energies of every mineral.

The relationship between Gibb's free energy F of a solid (or liquid) and confining pressure is:

(1) $$\left(\frac{\partial F}{\partial P}\right)_T = V \text{ where } V \text{ is mol volume.}$$

In other words a crystal exposed to rising pressure becomes increasingly unstable relative to a standard state not effected by pressure. The same condition may also be expressed in terms of chemical activity:

(2) $$\left(\frac{\partial(\ln a)}{\partial P}\right)_T = \frac{V}{RT}$$

If the effect of pressure on the free energy or chemical activity is compared with its effect on solubility in a fluid under same pressure as the crystal, a very significant difference is found. The following relationship is valid for saturated ideal solutions:

(3) $$\left(\frac{\partial(\ln c)}{\partial P}\right)_T = \frac{V_X - \overline{V}_L}{RT}$$

where c is saturated concentration, P is pressure on crystal and solution, V_X is molal volume of crystal, $\overline{V}_L$ partial molal volume of solute in solution, R gas constant, and T absolute temperature. V_X and $\overline{V}_L$ are closely equal in magnitude thus making the term $V_X - \overline{V}_L$ numerically small as compared with V_X (or $\overline{V}_L$). The effect of pressure on solubility is consequently much less than its effect on chemical activity according to equation (2). More significantly, there is no general tendency for $V_X - V_L$ to be positive, therefore, in contrast to free energy or activities, solubility does not generally increase with rising pressure, the opposite is just as likely. Herein lies a crucial difference between the «flow model» and the «diffusion model» of vein formation.

Let us then consider what happens when a dilation quartz-calcite vein form according to the «diffusion model». First we may discuss briefly the general opinion that open fractures do not form in deeper levels in the crust because such fractures mean increased overall volume and consequently increased potential energy at these high pressures. This argument seems valid for empty fractures at hydrostatic rock-pressure conditions but it does not necessarily hold for dilation veins in which minerals grow in pace with the dilation. The mineral matter which oozes into such veins may be derived from portions of the surrounding rocks where pressure is high. The volume shrinkage taking place by plastic adjustment in and around the source site could well more than compensate for the expansion of the fissure.

At anisotropic stress conditions it is even possible in terms of potential energy change, that empty tension fractures form and exist for some time at relatively high confining pressure. Under conditions of plastic flow and fracturing, for example, expansion in direction of least compresive stress partly by means of open tension gashes corresponds to a compression in the direction of maximum compressive stress. If the two stresses are sufficiently different in magnitude deformation with tension gashes could well represent a more energetically favorable situation than a nondeformed nonfractured rock complex even at considerable depth in the crust. Alternating layers of incompetent and competent rocks containing boudins may be examples on this situation. It seems thus that one cannot discard the possibility that open tension gashes form at deeper levels in the earth's crust. As pointed out above, however, field experience shows that open fractures are very rare in regional metamorphic terrains. On the other hand mineral filled fractures with walls separated up to several meters are common in metamorphic terrains, but as mentioned such dilation does not generally correspond to an overall volume increase if the source of the vein material is incorporated in the system under consideration.

There is evidence that the studied dilation-fracture veins grew in pace with the dilation and the fractures therefore at every stage of their evolution were practically filled with minerals. The other possibility, *viz.* that the fractures once were empty or filled with a fluid seems inconceivable because: 1) Presence of floating host-rock inclusions in many veins. 2) Some veins with comb structure have wall-to-wall crystals, *i. e.* single crystals of quartz or epidote which extend across the entire width of the vein, Pl. 2, Fig. E. In some boudinage structures the incompetent slate layers show no tendency to bend into the veins (fig. 2). 3) Empty cavities are exceedingly rare in the veins, and when present such cavities constitute but a minute portion of the whole vein.

The significance of point 2 above may perhaps not be obvious. The reasoning is as follows: If a vein grows in a fluid-filled fracture mineral nuclea would be expected to form along the wall and grow out from there to meet in the central portion of the vein. It is difficult to see how a single crystal could grow clear across from wall to wall because it would sooner or later collide with crystals growing in the opposite direction. If the fracture expands in pace with the growing crystals, however, it is conceivable that crystal growth takes place by adding matter to the crystals along their contact with the wall resulting in single wall-to-wall grains.

After these remarks we shall continue to follow what happens in a rock in which a tension fracture forms. Into this incipient minute fracture the most mobile constitutents of the surrounding rocks diffuse because of the free-energy inequalities caused by the fracturing. These constituents are probably mainly H_2O and CO_2 which exist as adsorbed and gaseous particles along the intergranular network in the rocks. Because of the high mobility of these volatile constituents a state of partial diffusion equilibrium will soon be attained at which the chemical potential of the most mobile constituents assume the same value in fracture as in surrounding rocks. If the fluid is gaseous the total pressure in the fracture equals the sum of the partial pressures of the constituents. The total fluid pressure in the fracture must generally be less than the rock pressure otherwise the rock would «boil» or sublimate causing smaller or larger pockets of fluids to form in the bulk of the rock. This is not what we find in metamorphic regions.

The fluid developing in the incipient minute fracture is liquid if the free energy of H_2O in the intergranular film of the rock is the same as — or higher than — the free energy of liquid water at the prevailing temperature. This does not mean that liquid water occurs in the compact bulk of the rock where pressure is high for the free energy of any liquid increases with rising pressure on the fluid, consequently the water energy in the system can very well be insufficient to develope liquid water in the rock and yet be sufficient to form liquid water in a fracture.

In addition to the highly fugitive constituents such as H_2O and CO_2, less mobile particles will diffuse toward the incipient fracture at a slower rate, the relative concentration in the fracture being controlled by relative mobility and availability. Prior to mineral nucleation in the fracture, when it still is very narrow, the majority of introduced particles are probably adsorbed along the fracture walls.

Diffusion must continue toward the fracture as long as the free energy of mobile particles here is lower than in the surrounding rocks. As the concentration of introduced mobile particles increases, their energies in the fracture also increase and sooner or later a stage will be reached at which vein minerals nucleate. Because of the low compressive stress in the fracture, nucleation and growth occur at a lower energy than that of the corresponding minerals in the compact part of the surrounding rocks. As long as tectonic forces tend to widen the fracture, thereby maintaining low compressive stress at this site, minerals will go on growing in the vein and diffusion continue toward it. Not until the tectonic stresses cease to operate do the vein minerals become subject to the full rock pressure. Their free energy consequently rises to the same (or even higher) level as that in the surroundings, and diffusion toward the vein must come to a halt. It is interesting to note that at this mature stage of the vein evolution a diffusion out from the vein is theoretically possible, because minerals grown at the low compressive stress in the vein may well become chemically incompatible with the surrounding-rock minerals when pressure is equalized between vein and enclosing rocks. Hydrous minerals, for example, are apt to form in fractures even in rather dry rocks because of the relative high concentration of water diffusing into fractures and because the partial water activity (vapor pressure) of hydrous minerals in fractures is relatively low. When ultimately the vein becomes effected by the full rock pressure the partial water activity of the hydrous vein minerals rises and could readily reach a level beyond that in the surroundings. The consequent reversed diffusion of water out from the vein is apt to form an external contact zone of hydrous minerals. Similar arguments can be used for other chemical constituents of vein minerals.

The above analyses of vein formation shows that diffusion toward dilating fractures and growth of minerals at such sites is not only possible, but indeed necessary for constituents which are mobile in rocks.

We have emphasized earlier that field observations indicate a strong syncronization between expansion of fractures and growth of minerals in these. It follows naturally from the theory described above that minerals should not tend to fill a tension fracture at (much) greater rate than the rate of widening as caused by tectonic forces. A forcefull growth of vein minerals would mean higher compressive stress on the vein minerals than in the surroundings and a consequent halt of diffusion and growth. On the other hand it is not obvious why filling of tension fractures do not generally take place at a slower rate than the tectonic dilation so that empty gashes would occur rather commonly in metamorphic rocks. A reasonable explanation of the apparent lack of lag between dilation and mineral filling is that tectonic stresses under regional metamorphism are generally incapable of causing tension gashes without co-operation from a medium which fills the gashes and thus helps to separate the fracture walls. Diabase magma pouring into crustal fractures, incompetent rocks flowing into cracks, or vein minerals growing in tension gashes may all represent the medium necessary to expand tension fractures in the crust. Such dilation aids are found to be necessary for formation of dilated tension fractures in some experiments dealing with deformation under anisotropic compressive stresses at high confining pressures. BRIDGMAN, (1952) who has investigated the behavior of various solids under anisotropic compressive stresses showed that under his experimental conditions, tension fractures would widen in the direction of least compressive stress only if the surrounding medium (oil) was permitted to flow into the fractures. The intruded oil would execute dilating force on the fracture walls somewhat larger in magnitude than the oppositly directed compressive stress which acted on the outer sides of the fractured solid. If the enclosing oil was hindered to enter incipient tension fractures (by means of a sufficiently rigid jacket) the fractures would of course not widen as long as all principal stresses were compressive.

Incidentally, Bridgman's experiments and his interpretations show that deformation under true tension and under anisotropic compression are not identical pheomena as sometimes assumed in geological discussion. Under true tensile stress fractures can widen without the aid of a filling substance.

Similar conditions may well exist during regional metamorphism, particularily at low levels in the crust: tension fractures may not widen unless they become filled with matter which creates enough pressure to overcome the component of compressive stress perpendicular to the fracture. In the studied veins the filling matter is the vein minerals which help to expand the fracture by their growth.

It is worth noting that the conditions under boudinage development are somewhat different from the Bridgman's experiments in an essential point. Contrary to the behavior of the enclosing oil in the experiments the incompetent rocks adjacent to a boudin layer become plastically elongate in one or more directions parallel to the boudin layer. This induces true tensile stress in the more rigid boudin layer, and it is conceivable that tension fractures would widen even if matter did not fill the gashes. Under the circumstances it is difficult to see how one might decide whether the filling (of vein minerals as well as plastically intruded adjacent rock) is necessary for the dilation or not. In any event it is obvious that the filling of the fractures aids their spatial dilation in terms of accellerated rate and greater ultimate width even if not always being necessary for the expansion.

COMPOSITIONAL RELATIONSHIP BETWEEN VEIN AND HOST ROCK

It has been pointed out above that the mineral assemblages in the studied veins are somewhat dependent on composition of host rock. We shall sum up these relationships briefly:

Albitic feldspar is common in veins in greenstone, it is rare in veins in graywacke and impure sandstone, and it is not found at all as vein mineral in argillaceous schists. The vein albite in greenstones is white, gray or greenish (due to chlorite inclusions); in graywacke and sandstone the vein albite is always reddish.

Chlorite is much more abundant in veins in greenstones than in argillaceous schists, besides the vein chlorite is generally more Fe-rich in greenstones than in the schists.

Biotite is rather common in veins in some greenstones, but exceedingly rare in veins in schists, graywacke and sandstone.

Actinolite is found in two veins in greenstone (Nos. 36418 and 36520).

Vein epidote is also a rare mineral encountered in but 5 veins in greenstones (Nos. 36519, 36515, 36417, 36449, 38500) and in epidote-rich inclusions in sandstone.

Other minor vein minerals such as tourmaline and some ore minerals occur too infrequently to justify conclusion as to their dependance on host rocks.

Quartz and calcite, the main vein minerals, occur in veins in all kinds of host rocks.

In spite of this compositional dependence of veins on host rock their evolution according to the model adopted in this paper represents a pronounced chemical differentiation inasmuch as there is much more quartz and/or calcite in the veins than in most country rocks. The fact that quartz is strongly enriched in most veins is understandable in view of the great chemical mobility which silica generally has during regional metamorphism. The same argument probably holds for calcite—its constituents CaO and CO_2 are mobile under regional metamorphism resulting in frequent occurence of calcite in dilation veins.

It is interesting to note that the chief minerals in veins in greenstones are related to the major host-rock minerals by a simple chemical reaction:

$$(4)\quad \underset{\text{epidote}}{2\,Ca_2Al_3Si_3O_{12}OH} + \underset{\text{actinolite}}{3\,Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2} + 10CO_2 + 8H_2O$$

$$= \underset{\text{chlorite}}{3\,(Mg,Fe)_5Al_2Si_3O_{10}(OH)_8} + \underset{\text{calcite}}{10\,CaCO_3} + \underset{\text{quartz}}{21\,SiO_2}$$

where the minerals of the left side with but few exceptions are confined to the host greenstones and those on the right side are the chief vein minerals.

Our model of vein formation is based upon a stress or pressure difference between host rock and vein. In consistence with that model we shall now investigate the possibility that the reaction above goes toward the right-hand side at the low total pressure in veins more readily than it does at the high pressure in the host rocks. Reaction (4) is somewhat inconvenient to study from this point of view because of its complexity. We shall therefore consider the simple wollastonite reaction:

$$(5)\qquad CaSiO_3 + CO_2 = CaCO_3 + SiO_2$$

in order to discuss the conditions at which carbonatization is limited to low-pressure fractures or «pressure shadows» in general. Let the chemically stable host rock consists of wollastonite and quartz. Temperature is uniform $= T$ throughout host rock and its fractures or veins. Host-rock pressure is P which for simplicity is assumed isotropic. Total pressure in a given fracture is p. The two pressures are unequal: $P > p$. The fracture expands continously by tectonic forces (+ aid from mineral fillings, see above) in such a manner that p remains less than P for some time. CO_2 exists as adsorbed molecules (complex ions) in the intergranular system in the host rock and in mosaic fractures or other fault structures in wollastonite and quartz. We assume that CO_2 is able to diffuse rather easily through the rock and its fractures and that there is a supply of CO_2 outside the rock complex. If calcite or other carbonates are stabilized within the rock complex their growth is therefore not prevented by lack of CO_2.

Since calcite does not exist in the stable wollastonite fels, the partial free energy of CO_2 in the rock is too low to stabilize calcite at the high rock pressure. The question is now: is it feasible that calcite be stable at the low pressure (stress) in fractures and «pressure shadows» at a free energy of CO_2 which is insufficient to stabilize calcite in the host rock? If this is true the CO_2 molecules needed to form calcite in an expanding fracture can diffuse through the surrounding wollastonite fels without causing the slightest trace of carbonatization. The question can be answered most rigorously if one considers the situation in terms of thermodynamics. Let the Gibbs free energies of the constituent oxides in the rock be: $(F_q)_{P,T}$, $(\overline{F}^{\text{woll}}_{\text{CaO}})_{P,T}$ and F_{CO_2} where subscript q refers to quartz and subscript P, T indicates that the energies refer to given P, T conditions. $\overline{F}^{\text{woll}}_{\text{CaO}}$ is the partial free energy of CaO in wollastonite.

We shall recall at this point that only the total free energy of any given stoechiometric compound — such as wollastonite — is invariant at given pressure and temperature. The partial free energies can vary at constant P and T, the one dependent on the others. For wollastonite at P and T, for example, the relation between total and partial free energies is:

$$(6) \qquad (F^{\text{woll}})_{P,T} = \text{constant} = (\overline{F}^{\text{woll}}_{\text{CaO}})_{P,T} + (\overline{F}^{\text{woll}}_{\text{SiO}_2})_{P,T}$$

where $(F_{\text{woll}})_{P,T}$ is constant but $(\overline{F}^{\text{woll}}_{\text{CaO}})_{P,T}$ and $(\overline{F}^{\text{woll}}_{\text{SiO}_2})_{P,T}$ can vary depending upon type of minerals present.

(The mass action law states that the product of the activities is constant at given P, T conditions. Since partial free energy is proportional to the logarithm of the activities, it follows that the sum of the partial free energies is constant at given P, T values).

In the stable wollastonite-quartz fels the partial free energy of silica in wollastonite, $(\bar{F}^{\text{woll}}_{\text{SiO}_2})_{\text{P.T}}$ must equal the energy of coexisting quartz: $(\bar{F}^{\text{woll}}_{\text{SiO}_2})_{\text{P,T}} = (F_{\text{q}})_{\text{P,T}}$ because the two minerals are in equilibrium with each other. $(\bar{F}^{\text{woll}}_{\text{CaO}})_{P,T}$ is then given by the mass action law:

$$(7) \qquad (\bar{F}^{\text{woll}}_{\text{Cao}})_{\text{P,T}} = (F^{\text{woll}})_{\text{P,T}} - (F_{\text{q}})_{\text{P,T}}$$

(If now quartz was replaced by another mineral with different free energy of SiO$_2$, say $(\bar{F}^{X}_{SiO_2})_{P,T}$, then the free energy of CaO in wollastonite must change to a value $(\bar{F}^{\text{woll}}_{\text{CaO}})'_{\text{P,T}}$ in order to conform to the new condition:

$$(8) \qquad (F^{\text{woll}})_{\text{P,T}} = \text{const.} = (\bar{F}^{\text{woll}}_{\text{CaO}})'_{\text{P,T}} + (\bar{F}^{\text{X}}_{\text{SiO}_2})_{\text{P,T}}).$$

Since calcite is unstable in the host rock the sum of $(\bar{F}^{\text{woll}}_{\text{CaO}})_{\text{P,T}}$ and F_{CO_2} is less than the total free energy of calcite at the given P and T:

$$(9) \qquad (\bar{F}^{\text{woll}}_{\text{CaO}})_{\text{P,T}} + \bar{F}_{\text{CO}_2} < (F^{\text{cal}})_{\text{P.T}} = \text{constant.}$$

If $(\bar{F}^{\text{woll}}_{\text{CaO}})_{\text{P.T}} + \bar{F}_{\text{CO}_2} \geqq (F^{\text{cal}})_{\text{P,T}}$ calcite would form by reaction between wollastonite and CO$_2$ in the rock.

Let us then consider the conditions in an expanding fracture which is fed by diffusion from the surroundings and in which the total pressure is p. The free-energy values of the various constituents here can clearly not be greater than those in the bulk of the rock because if they were, diffusion toward the fracture would cease. Hence minerals can grow in the fracture only if their energies are lower than $(F_{\text{q}})_{\text{P,T}}$, $(\bar{F}^{\text{woll}}_{\text{CaO}})$ and F_{CO_2} which are the free energies in the host rock. The observed fact that quartz grows in fractures puts a limit on the free energy of silica at these sites if we permit ourselves to assume that the growth of quartz is so slow that supersaturation can be neglected. Hence the free energy of silica in the fracture is:

$$(10) \qquad (F_{\text{SiO}_2})_{p,\text{T}} = (F_{\text{q}})_{p,\text{T}} = (F_{\text{q}})_{\text{P,T}} - (P - p)V_{\text{q}}$$

where the last term represents the decrement in free energy because of the pressure difference between host rock and vein.

The high mobility of CO$_2$ equalizes rapidly any gradient in F_{CO_2} which may occur. We can therefore consider the free energy of CO$_2$ in fractures to be same as that in the host *i. e.* $(F_{\text{CO}_2})_{p,\text{T}} = (F_{\text{CO}_2})_{\text{P,T}} = F_{\text{CO}_2}$. (Note here that P and p do not refer to the partial pressure of CO$_2$, but rather to rock- and fracture environment respectively.) Under these circumstances the free energy of CaO required to stabilize calcite and wollastonite respectively, in the fracture are:

(11) $$(\bar{F}^{\mathrm{cal}}_{\mathrm{CaO}})_{\mathrm{p,T}} = (F^{\mathrm{cal}})_{\mathrm{p,T}} - (F_{\mathrm{CO_2}})_{\mathrm{p,T}} = (\bar{F}^{\mathrm{woll}}_{\mathrm{CaO}})_{\mathrm{P,T}} + (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} - (P - p)\, V_{\mathrm{cal}} - F_{\mathrm{CO_2}},$$

and:

(12) $$(\bar{F}^{\mathrm{woll}}_{\mathrm{CaO}})_{\mathrm{p,T}} = (F^{\mathrm{woll}})_{\mathrm{p,T}} - (F_{\mathrm{q}})_{\mathrm{p,T}} = (\bar{F}^{\mathrm{woll}}_{\mathrm{CaO}})_{\mathrm{P,T}} + (F_{\mathrm{q}})_{\mathrm{P,T}} - (P - p)\, V_{\mathrm{woll}} - (F_{\mathrm{q}})_{\mathrm{P,T}} + (P - p) V_{\mathrm{q}}$$

where $(\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}}$ is the energy of CO_2 necessary to stabilize calcite at the high pressure in the bulk of the rock. Calcite will form when

(13) $$(\bar{F}^{\mathrm{cal}}_{\mathrm{CaO}})_{\mathrm{p,T}} < (\bar{F}^{\mathrm{woll}}_{\mathrm{CaO}})_{\mathrm{p,T}},\ \text{i. e.}$$

when:

(14) $$(\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} - F_{\mathrm{CO_2}}, < (P - p)\, (V_{\mathrm{cal}} + V_{\mathrm{q}} - V_{\mathrm{woll}})$$

Rearranging eq. (14) gives condition for formation of calcite in the fracture in terms of the partial CO_2-energy, $F_{\mathrm{CO_2}}$ of the system:

(15) $$F_{\mathrm{CO_2}} > (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} + (P - p)\, (V_{\mathrm{woll}} - V_{\mathrm{q}} - V_{\mathrm{cal}})$$

$V_{\mathrm{woll}} = 41$ cc. $V_{\mathrm{q}} = 22$ cc, $V_{\mathrm{cal}} = 37$ cc, hence calcite is stabilized in a fracture when the following unequality is satisfied:

(16) $$F_{\mathrm{CO_2}} > (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} - (P - p)\ 18\ \text{ccatm.}$$

If $F_{\mathrm{CO_2}}$ in the rock and fracture equals $(\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}}$, calcite is not only stabilized in the fracture but also in the bulk of the rock. The occurrence of calcite is hence determined by the following conditions:

If $F_{\mathrm{CO_2}} < (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} - (P - p)$ 18 ccatm. calcite can not form stably either in a fraction or in the rock.

If $(\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} > F_{\mathrm{CO_2}} > (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}} - (P - p)$ 18 ccatm. calcite can form stably in fractures but not in the bulk of rock.

If $F_{\mathrm{CO_2}} \geqq (\bar{F}^{\mathrm{cal}}_{\mathrm{CO_2}})_{\mathrm{P,T}}$ calcite is stable in fracture as well as in the host rock.

Arguments analogous to those put forth above can be applied to hydration reactions showing that it is thermodynamically possible for hydrous minerals to develop in fractures or other low-pressure sites at a free energy of water which is insufficient to stabilize the same hydrous minerals in the host rock.

In principle the effect of stress or pressure differences on the wollastonite-calcite reaction and simple hydration reactions is not different from the effect of stress or pressure differences on complex reactions examplified by eq. (4).

In view of the discussion above the writer concludes that calcite, quartz and some chlorite grew in the studied veins supplied by CO_2, H_2O, SiO_2, CaO and lesser amounts of MgO, FeO, Al_2O_3 which migrated through and from the surrounding rocks. The source of CaO, FeO, MgO, Al_2O_3 and SiO_2 is probably the not-too-distant host rocks (greenstones). H_2O and CO_2 may have come from more distant sources. Incidentally, reactions between clay minerals and calcite in sediments in the area would be natural producers of CO_2 and H_2O.

Many of the studied greenstone host rocks contain carbonate beside actinolite, epidote etc. For these rocks and their calcite-quartz veins the ΔF_{CO_2} value is zero, *i. e.* the actual F_{CO_2} - value is the same as that necessary to stabilize calcite in the rock. According to the discussion above this is a condition at which actinolite and epidote should not be stable together in a fracture filling. The constituents of these silicates should react with CO_2 in a fracture and give calcite, quartz and chlorite according to reaction (4). In this connection it is most interestinging to recall that actinolite is found only in small amounts in two of the several hundred veins studied, and epidote in but five occurrences. Epidote and actinolite are not found together in any of the veins. On the other hand chlorite is present in most of the veins in greenstones which themselves contain actinolite and epidote among the major minerals.

The presence of biotite in the veins deserves some consideration. Biotite is much less common in veins than chlorite, but it occurs in a number of veins in greenstone. It is very rare as vein mineral in the metasediments. This seems puzzling in view of the fact that the sediments are generally richer in potash than the greenstone (Table 5), potash of course being a major constituent of biotite. An observation which seems significant in this connection is that biotite containing veins in greenstones are generally surrounded by a strongly bleached zone depleted in ferromagnesian minerals such as chlorite and actinolite. This observation, in addition to the fact that the greenstone are considerably richer in FeO, MgO and TiO_2 than the metasediments, (Table 5) suggests that Fe and Mg (TiO_2) control the formation of biotite rather than K. If the free energy of FeO, MgO and TiO_2 is higher in some greenstones than in the metasediments — a likely condition in view of the difference in mineralogy — higher F_{FeO}, F_{TiO_2} and F_{MgO} values will also develop in greenstone fractures, and correspondingly smaller value of F_{K_2O} is needed to develope biotite in such fractures. Even so, the F_{K_2O} -value in greenstone has been too low in most places to produce biotite in low-pressure veins, but occasionally the energy level of potash has been sufficient for this process. Stabilization of biotite at the high pressure in the host rock requires higher free energies of MgO, FeO and K_2O. Biotite is correspondingly rare in the bulk of the greenstones. If now potassium in one form or another was more mobile than iron, magnesium and aluminum, K_2O would play a similar role in biotite formation as CO_2 does in calcite formation. Potassium could diffuse toward the growing biotite in the vein from a large volume of the surrounding potash-poor rocks whereas iron and magnesium are depleted from a much more limited region around the vein such as seen in the field.

A similar explanation is probably valid for the occurrence of vein tourmaline. This mineral is found in two or three veins in greenstone, but not as vein constituent in the metasediments. Yet the most critical constituent of tourmaline, boron, is generally much more concentrated in sediments than in basic lavas (greenstones), see Goldschmidt 1932, Rankama, 1950. Analyses of some of the studied rocks show the same trend, table 6. The studied tourmaline containing veins are almost invariably surrounded by intensively bleached host

Table 6. Boron content in rocks.

	Greenstones			Serisitic sandstones			Slates	
	19566	35796	36430	36566	36573	36582	35741	36554
ppm	< 2	< 2	< 2	~ 10	< 2	~ 5	> 100	> 100

Analyst: S. Rutlin.

rocks and contain sometime bleached host-rock inclusions, see table 1. It seems obvious that iron and magnesium in tourmaline has been extracted from the host rocks. A high energy level of Fe and Mg in greenstone could cause tourmaline to form in fractures in spite of the energy level of boron being less than in the metasediments. Supply of the mobile boron from a large region of surrounding rocks and depletion of the less mobile iron and magnesium from adjacent rocks would nourish further growth of tourmaline in the fracture vein. If boron exists mainly in small concentration as adsorbed particles in the intergranular of the greenstone it would be very difficult to detect concentration gradients around the vein. The source of boron is then either to be considered as dissiminated in very dilute concentration throughout the surrounding rocks, or perhaps as more lozalized boron-rich sites separated in space from the vein.

A more conventional explanation of the origin of biotite and tourmaline in vein with bleached external zones is that potassium and boron were introduced along the vein and biotite and tourmaline were formed by extracting iron, magnesium etc. from the adjacent rocks. A fact seemingly in favor of such a mechanism is lack of specific boron and potassium minerals in the host rock. In view of this it may be reasonable to assume that potassium and boron could not have moved through the host rocks. It has been documented above, however, that this objection is not generally valid: it is thermodynamically possible for a constituent which diffuses through a rock to build specific minerals in fractures and thus be enriched there without leaving trace of its passage through the bulk of the rock. Since this is a possible process, and the veins are small confined bodies without visible avenues of supply, the writer concludes that also the boron and potassium occurring in some veins in greenstones diffused through the bulk of the adjacent rocks.

Acknowledgment

The field work on which the present study was based was done while the writer recived economic support from The Council of General Science, Norway. Thanks are due to Dr. R. Kretz, Mr. S. Rutlin and Mr. O. Joensuu fot the spectroscopib analyses; and to Mr. J. McLelland for determination of refraction and calculation of analyses.

REFERENCES

BRIDGMAN, P. W., 1952: Studies in large plasticflow and fractures, McGraw-Hill, New York, pp. 1—362.

BATEMAN, A. M. 1951: The formation of mineral deposits, New York, pp. 1—371.

CAMERON, E. N., JAHNS, R. H., MCNAIR, A. H. PAGE, J. J. 1949: Internal structures of granitic pegmatites, Econ. Geol. Mon. 2. pp. 1—115.

FAIRBAIRN, H. W. 1941: Deformation lamellae in quartz etc. Bull. Geol. Soc. Am. Vol. 52 pp. 1265—1278.

— 1950: Synthetic quartzite, Amer. Min. 35. p. 735.

FOLK, R. L. and WEAVER, C. E. 1952: A study of the texture and composition of chert, Am. Jour. Sci. Vol. 250, pp. 498—510.

INGERSON, E., and TUTTLE, O. F., 1945: Relations of lamellae and crystallography of quartz and fabric directions in some deformed rocks, Trans. Am. Heoph. Union, Vol. 26, pp. 95—106.

KUENEN, PH.H., 1938: Observations and experiments in ptygmatic folding, Comptes rendus Soc. geol. Finlande, No. 12, p.11.

PELTO, C. R., 1956: A study of chalcedony, Am. Jour. Sci. vol. 254, pp. 32—50.

RAMBERG, H., 1945: Thermodynamics of the earth's crust, I Norsk geol. Tidsskr. vol. 24, pp. 98—111.

— 1952: Origin of metamorphic and metasomatic Rocks, Univ. of Chicago Press, pp. 1—317.

— 1956: Pegmatites in West Greenland, Bull. Geol. Soc. Am. Vol. 67, pp. 185—214.

— 1959: Evolution of ptygmatic folding, Norsk geol. Tidsskr., vol. 39, pp. 99—152.

— 1961: Artificial and natural photoelastic effects in quartz and feldspar, Am.min., in press.

REITAN, P., 1958: Pegmatite Veins and the surrounding Rocks. II. Changes in the olivine gabbro surrounding three pegmatite veins, Risør, Norway, Norsk geol. Tidsskr., vol. 38. pp. 279—311

— 1959: Pegmatite Veins and the surrounding Rocks. III. Structural control of small pegmatites in amphibolite, Rytterholmen, Kragerøfjord, Norway. — IV. Genesis of a discordant pegmatite vein, St.Hansholmen, Risør, Norway, Norsk geol. Tidsskr., vol. 39, pp. 175—195; 197—229.

RIEDEL, W. 1929: Zur. Mechanik Geologischer Bruchercheinungen, Zentralbl. Min. 1, pp. 354—368.

GOLDSCHMIDT, V. M., and PETERS, Cl., 1932: Zur Geochemie des Bors, Nachr. Ges. Wiss. Gottingen, Mat.physik kl. 111, 28. IV, p. 528.

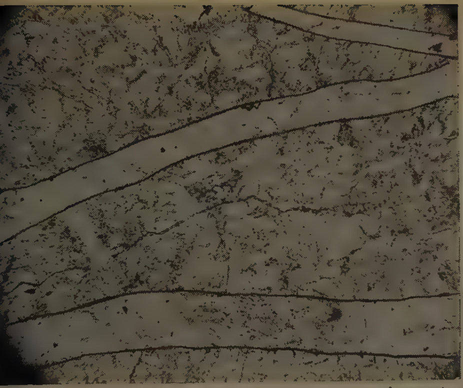
RANKAMA, K. and SAHAMA, TH. G., 1950: Geochemistry, Chicago.

Manuscript received May 15, 1959.

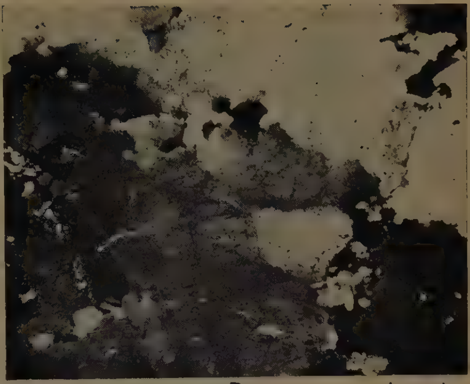
Printed March 1961.

Plate 1.

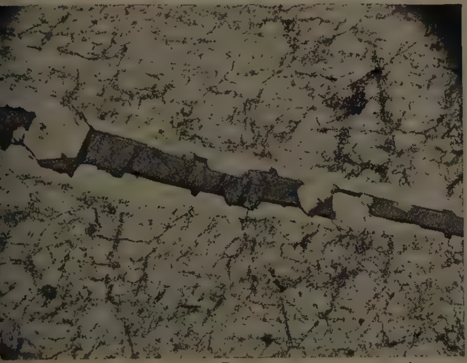
A. Clear phantom quartz veinlets in "dusty" quartz.
B. Same as A under crossed nicols.
C. Calcite veinlets in "dusty" quartz surrounded by clear quartz. Note small euhedral quartz crystals in calcite.
D. Veinlets crossing through greenstone (left part of picture) and albite aggregate. Note chlorite in center of vein where it cuts the greenstone. The dark color in the lower portion of the vein is due to "dusty" feldspar.
E. Stylolite seam in vein quartz.
F. Remnants of replaced schist in vein quartz.



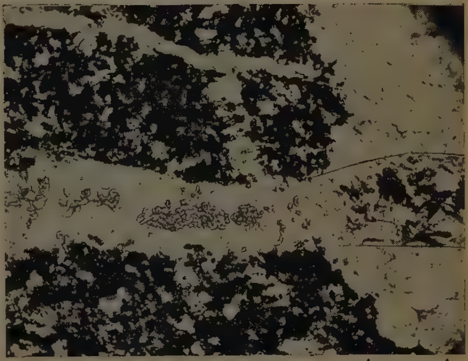
A



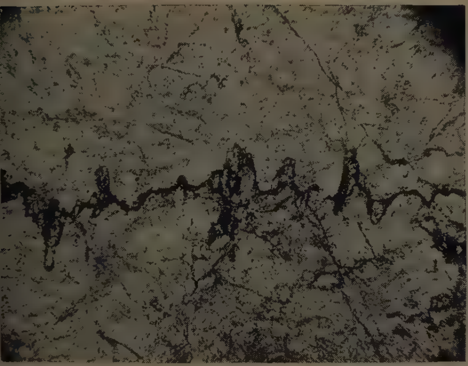
B



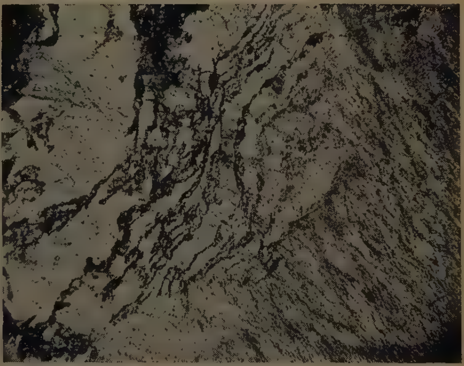
C



D



E

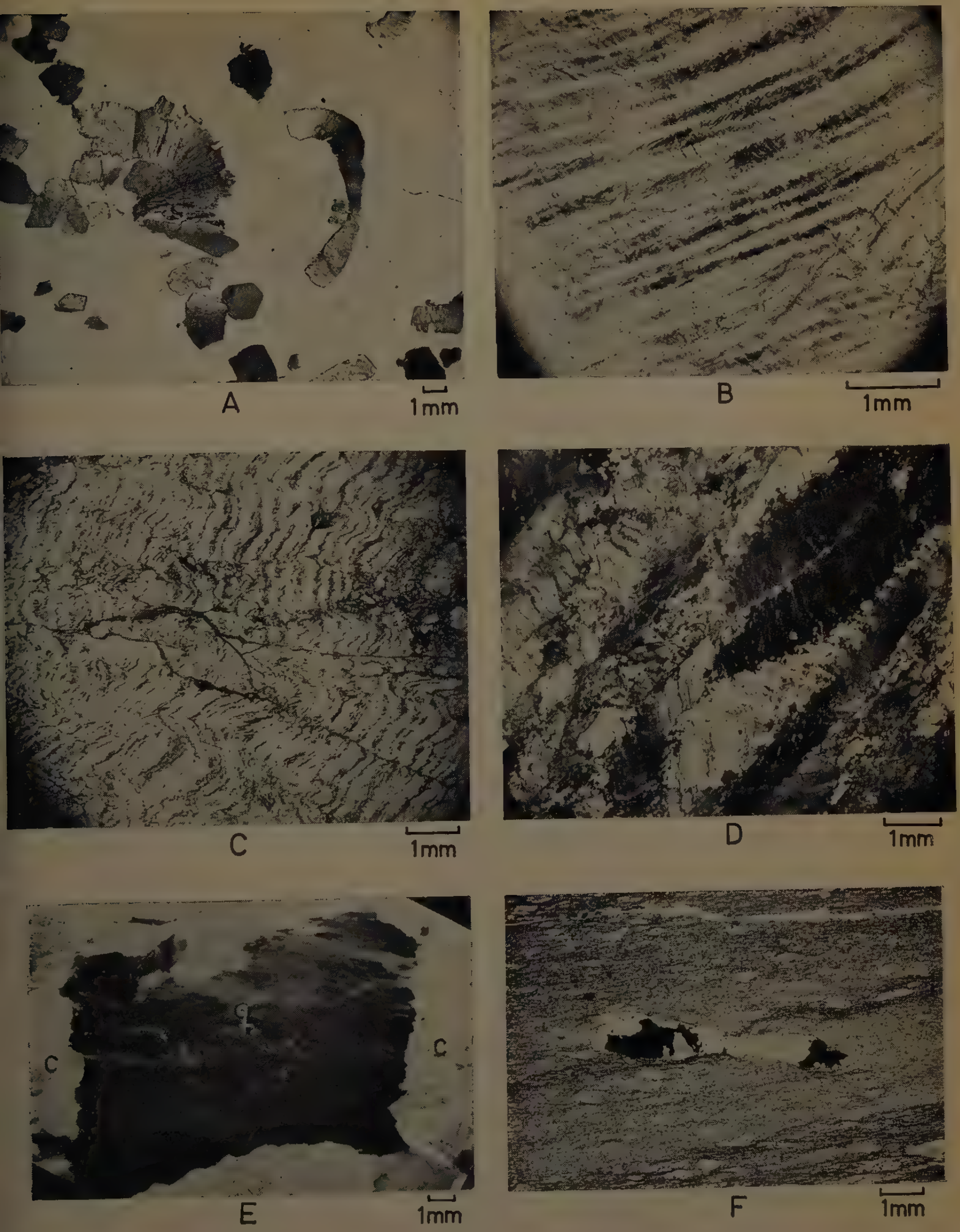


F

Plate 2.

A. Chlorite inclusions in vein quartz.
B. Cloudy lamellae in quartz.
C. Folded bands of sericite in vein quartz.
D. Same as C under crossed nicols. Note that picture is rotated about 45 ° anticlockwise relative to C.
E. Wall-to-wall quartz crystal in vein marble (C). Crossed nicols.
F. Quartz in pressure shadow between pyrite crystals in slate. The two pyrite grains (black) are probably fragments of one crystal fractured and separated in boudinage-like fashion.

PLATE 2.



ORDOVICIAN CLIMATIC ZONES

By

NILS SPJELDNÆS
(Paleontologisk Museum, Oslo NØ, Norway)

Abstract: It is suggested that the Ordovician faunal provinces largely represent climatic zones, and the biological and lithological argument for and against this hypotesis are discussed. The pattern of climatic zones obtained indicate that 1) that Eur-Africa and the Americas have drifted 30—40° apart in an E-W direction since the Ordovician, but the evidence for this is not conclusive. 2) that one of the poles was in or just W. of Central Africa. Both of these features are supported by paleomagnetic studies. It is suggested that polar ice-caps existed during the Ordovician, and that the close correlation of the climatic changes, with faunal migrations and transgressions-regression might be used for long distance correlations. The correlation between climatic changes and orogenies could be observed, but it is not very good, and there is no good correlation between climatic changes and volcanic activity in the Ordovician.

Introduction.

Not much has been done in the study of the climate of the Ordovician, even though a large number of details and isolated observations have been published. Due to this, and the difficulties in interpretation of the observations, most general papers on paleoclimatology have treated the Ordovician rather briefly, and generally (i.e. SCHWARTZBACH 1950) suggested that a warm and uniform climate prevailed during that period.

The present study started with the authors work on the Ordovician faunas of Europe and North Africa, the details of which might be published as a separate paper. The references to the vast litterature on the stratigraphy, lithology and regional distribution of the Ordovician is therefore ommitted here. The same is the case with the references to the litterature on the Ordovician orogenies and volcanic activity.

The results given in this paper have been presented at some occasions earlier (cf. SPJELDNÆS 1958), and the author is indebted to a number of scientists for valuable critizism, and advice. Among them, I would like to mention professor K. C. Caster, Cincinnati, Dr. P. C. Cloud jr. and. Dr. A. R. Palmer of the U.S. Geological Survey, and professor O. Holtedahl, Oslo. I am especially indebted to professor W. C. Sweet, Ohio State University, who has not only discussed the problems, but also read the manuscript critically.

I wish also to express my thanks to Mrs. K. Gran and Miss I. Lowzow who have made the maps and diagrams.

Faunal Provinces.

Not knowing the exact ecology of the Ordovician animals, it is very difficult to limit faunal provinces which will equal the recent zoogeographical ones. The faunas will of course vary widely with depth and oxygen content of the water, and the present study is therefore restricted to the non-euxinic, fairly shallow water shell deposits. Fortunately these deposists were widespread in the Ordovician, probably more so than in any other period. The faunal elements considered are chiefly the benthonic and semibenthonic. Plancthonic forms, mainly graptolites, are not considered here. The primary factor regulating the distribution of the faunal provinces is regarded to be the water temperature.

As mentioned below, the borders of the faunal provinces change with time, and like the recent faunal provinces or zones, they are believed to be controlled by the climate. The changes in their distribution is regarded to be due to climatic changes.

The lithology of the sediments in which the faunas are found, reflects the ecology of the area, and also, to a certain degree, the temperature of the water. The characteristic lithologic features of each faunal province that are supposed to be of some climatological importance, are therefore mentioned in the description of the faunal provinces.

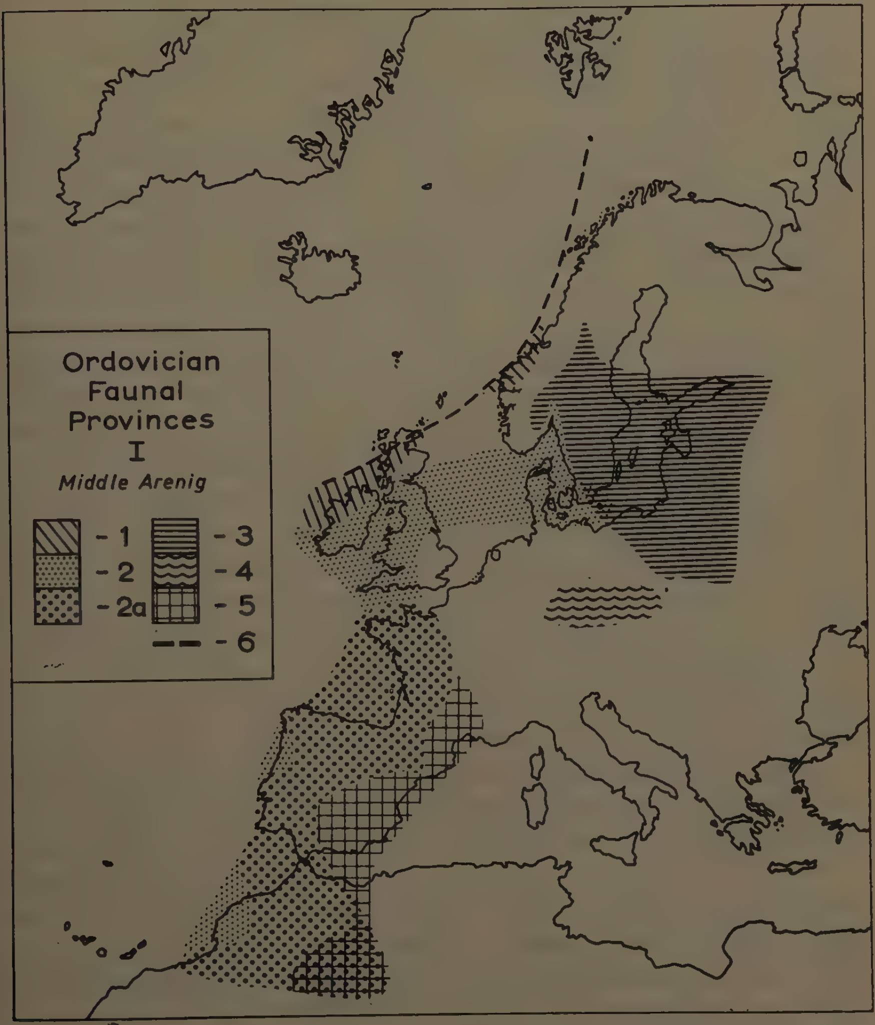


Fig. 1. Ordovician Faunal Provinces in Europe in the middle Arenig. 1 = Faunas of Scoto-Appalachian type, 2 = graptolite shale facies (both in Anglo-Scandic and Mediterranean province), 3 = Shallow water limestone facies of the Scando-Baltic (Orthoceros-limestone facies 2a = Armorican Quartzite, 4 = Bohemian facies, 5 = Trilobite facies, mudrock in Mediterranean province. 6 = S-E-ern limit of American-Arctic province.

THE AMERICAN-ARCTIC PROVINCE.

This province is, in its typical development, found in the N-W part of Europe: in Scotland (Durness), Norway (Smøla), Bear Island and W Spitsbergen. It has its main distribution in Greenland, Arctic and Central Canada, and in the U.S.A.

In the Lower Ordovician (Canadian) the American-Arctic province included large parts of North America, but in the Middle Ordovician it was resticted to a belt extending from Greenland through Arctic Canada, the Rocky Mountains region, to New Mexico and western Texas.

The fauna is thick-shelled, and is dominated by mollusks, especially gastropods and cephalopods. Sponges and corals are also common, end reefs are found from the base of the Middle Ordovician. The rocks in which the fossils are found are mainly massive limestones and dolomites. The sandstones and shales found are generally of the orthoquartzite suite, in this case probably indicating an origin by rapid and complete weathering in a warm climate.

The typical fauna of this province is the upper Middle Ordovician Red River fauna, especially rich in cephalopods, which according to FLOWER (1957) indicate warm water.

In Europe this faunal province is mainly represented in the Lower Ordovician. Only in Durness and Bear Island have Middle Ordovician beds been positively identified. The Bear Island beds were correlated with the Black River by HOLTEDAHL (1919), on the base of the cephalopods. The best preserved forms, *Gonioceras* sp. and *Actinoceras ursium* FOERSTE (1921), indicate that this is correct. *Gonioceras* ranges through the whole Middle Ordovician, and *Actinoceras ursium* belongs to the group of *A. centrale*, which according to FLOWER (1957, pp. 33—34), ranges from the Lowville through the Rockland in North America, probably being a good indicator of Wilderness (=Black River) age.

The area in which the beds of this faunal province were deposited was a stable shelf. This fauna is generally not found in areas with contemporaneous folding and volcanism, and the «geosynclines» were generally shallow depressions, in which no orogenic movement took place, at least not until long after the Ordovician. One of the few exceptions is the Smøla locality, where beds are found in an area of Ordovician orogeny and volcanism. It has to be mentioned that the Lower Ordovician (Canadian) part of the sequence in Europe (Durness, Smøla, Bear Island and Spitzbergen) might belong to either this province, or to the Scoto-Appalaccic. The development of the upper part of the Lower Ordovician is fairly similar lithologically in these two provinces, and the faunas found in Europe are small and not conclusive. The Smøla fauna might also be Middle Ordovician, not Lower.

The shelf deposits of this province are limited by graptolite shale and siliceous rocks in the western part of North America (Vinini Formation). In the south-east (Oklahoma) the same conditions are found. In the typical area the shelf deposits are rather incomplete The Lower Ordovician is represented by dolomitic rocks (Manitou a.o. form.) which are followed by high Middle Ordovician sandstones (Harding, Eureka a.o.). The youngest member of the Ordovician is the generally coarse-grained dolomitic rocks with the typical Red River Fauna. (Fremont, Bighorn, Hanson Creek a.o.). Along the edges of the geosynclines the sequence is more complete. The low Middle Ordovician Whiterock stage is found in Nevada, Western Utah and Oklahoma. ("Upper Pogonip", Joins and Oil Creek Formations).

The Whiterock Stage is known from Greenland and possibly Durness (Poulsen 1951). In continental Europe, it is found in the Hølonda Limestone Trondheim Area, Norway, where *Goniotelus*, *Anomalorthis* and *Syndielasma* are characteristic representatives. At Chabarova, Northern Russia, material collected by Nansen has yielded an Appalachian fauna including *Isoteloides* and a genus resembling *Paraplethopeltis*.

The rocks in which this fauna is found are generally highly calcareous, but also contain amounts of shale, and the fauna is generally rather thin-shelled, in spite of some few corals, sponges and thick-shelled forms present. The forms mentioned above are typical, others are described by Ross (1951) and Hintze (1952).

It is possible that the higher Middle Ordovician in the Central Basin of Tennesse, in Kentucy, the Cinncinnati region and the Upper Mississippi Valley belong to this faunal province. It might also be a transition between this and the following province.

The Table Head beds in New Foundland and the Tourmakeady beds in Western Eire might also belong to the same faunal province, but they were deposited under different conditions, as geosynclinal sediments. In Eire they are also interbedded with volcanic rocks. They are, therefore, not considered further here.

THE APPALACHIAN AND ANGLO-SCANDIC PROVINCES.

These two provinces, which are intimately related formed a broad belt trending from Alabama along the eastern part of North America, and continuing along the North-Western part of Europe, through Great Britain, Eire, southern Scandinavia and Esthonia. The provinces crossed the Caledonian system of geosynclines at a low angle. This produced a very complicated picture of the distribution of the faunas, partly because there were active geosynclines (subsiding areas), orogenic movements, volcanism and other features that led to rapid changes in environment, both horizontally and vertically (in time). Furthermore later orogenic events partly destroyed the beds and fossils through metamorphism, and partly displaced them geographically through folding and thrust-faulting.

The Appalchian province is found in Eastern N. America and the Girvan District of Scotland, including parts of Ireland. The Anglo-Scandic anclude parts of Eire and Wales, England, the Oslo Region in Norway, and the Baltic Region. In addition it may also be found in Belgium, and perhaps in the Carnic Alps. Both these provinces contains a complete range of sediments from euxinic graptolite shales with exclusively planctonic faunas, thick geosynclinal sediments, coarsegrained littoral or even terrestrial sediments to shallow-water limestones indicating oxygen-rich conditions with a rich benthonic fauna.

The faunas found in the different environment in these provinces are, of course, widely different, and this has led to great difficulties in correlation of the beds. A typical example of this is found in the Chasmops Series (Middle and Upper Middle Ordovician) of the Oslo Region (STØRMER 1953). Along the western and northern margin of the Region, there are beds which indicate deposition in very shallow water (calcarenites, constant silt content in terrigenous material, rich algal flora). The fauna is very similar to that in Esthonia, which is found in beds deposited on a stable shelf, under the same conditions. Nearer the central part of the region (Ringerike, parts of Hadeland), there are beds which were also deposited in ventilated but slightly deeper water, (absence of calcareous algae, few calcarenites, little silt, mainly clay and marl sediments). The fauna is remarkably similar to that found in beds of same age in Shropshire, where at least some of the beds were deposited under the same condition as the Norwegian beds mentioned.

In the central part of the Oslo Region, the beds were deposited under partly euxinic conditions, with occasional ventilation, and the fauna is highly variable, both horizontally and vertically because of this. The rocks are mainly dark gray to black shales with some marly limestones. Bedded calcarenites and calcareous algae are unknown. Similar faunas are found in beds probably deposited under the same condition in North Wales.

It is generally easier to correlate the types of faunas mentioned from the Oslo Region with the resembling distant ones than to correlate the faunas within the Region (cf. SPJELDNÆS 1957, pp. 196—199).

This indicates that all these faunas, in spite of their differences, belong to one faunal province, because in the whole area a similar environment would have produced a similar fauna. In each shallow clear-water environment, the baltic fauna would have appeared, in an environronment with changing oxygen content, the North Wales fauna would have appeared. This also indicates that the factor governing the distribution of the faunal provinces was temperature (=climate).

Because of the complexity of the faunal provinces mentioned here, it is very difficult to define them. There is also a gradual transition between them, and a certain amount of interfingering. In the Oslo Region the number of Appalachian forms is higher in the north than in the south (in the same environment). There is more similarity between the fauna of the Oslo Region and the Girvan district than between the Girvan faunas and those in England. In the Appalachian region, there are certain formations which have a much more «Baltic» fauna than these above and below (Boutetort Lst., Oranda and, especially, Pratts Ferry), indicating that they are "fingers" with Anglo-Scandic faunas.

Generally the Anglo-Scandic elements in North America, and the Appalachian ones in Europe (except Scotland and Ireland) are a small fraction of the species present. An exception from this rule is the fauna described by COOPER & KINDLE (1936) from Percé, Canada, which is dominantly Anglo-Scandic.

The resemblance is generally at the generic level. The percentage of common genera and species might be used to evaluate the relative positions of the continents in the Ordovician (see below).

It is difficult to define characteristic genera and species for the two provinces, for the genera that are common anough to define the faunas are generally restricted to one of the several environments, or they are found in both provinces, but often with a different vertical distribution. The genus *Christiania*, f.i. is known from the basal Middle Ordovician in Esthonia, the Leningrad District, and the Oslo Region, continues intermittently through the whole Ordovician in the latter area, is restricted to the Porterfield Stage in the Appalachians, Girvan and Central Eire, and is found in the Ashgillian in England, Wales and Belgium. Similar patterns of distribution are found in other genera also. A number of genera are restricted to one province, but they are generally not common and widespread. One of the few sets of genera that seem to be characteristic includes *Calliops*, which is the dominating phacopid trilobite in the Scoto-Appalachian province and *Chasmops*, which is the common one in the Anglo-Scandic province.

Further studies of the faunas in the two provinces will certainly lead to a higher number of common and characteristic genera or even species.

The Southern and Eastern limits of the Anglo-Scandic province are not well defined, it is found in the Caradocian of Belgium, and some of the faunal elements, otherwise unknown in the Mediterranean province, are found in the Carnic Alps (*Christiania* and *Sowerbyella*). The faunas typical of this province are also found in Kazakstan (RUKAVIŠNIKOVA 1956 a.o.) and possibly in Himalaya and in Western S. America.

The papers by TRUMPY (1943), and especially by HARRINGTON (in JENKS 1956) and HARRINGTON & LEANZA (1957) show that the Lower Ordovician of South America is strikingly similar to that of Northern Europe (Scandinavia and Great Britain). The Middle Ordovician is less well developed, and it is generally found as graptolite shales. There is, however, a slight indication in the development of facies (cf. HARRINGTON in JENKS 1956) that the western parts of the region belong to the Anglo-Scandic faunal province, and some of the eastern ones to the Mediterranean one, with *Synhomalonotus*.¹)

---

¹) After this was printed, the author became aware that WHITTARD (Mem. Paleontogr. Soc. *139*, 1960) had established *Neseuretus* to be the senior synonym of *Synhomalonotus*.

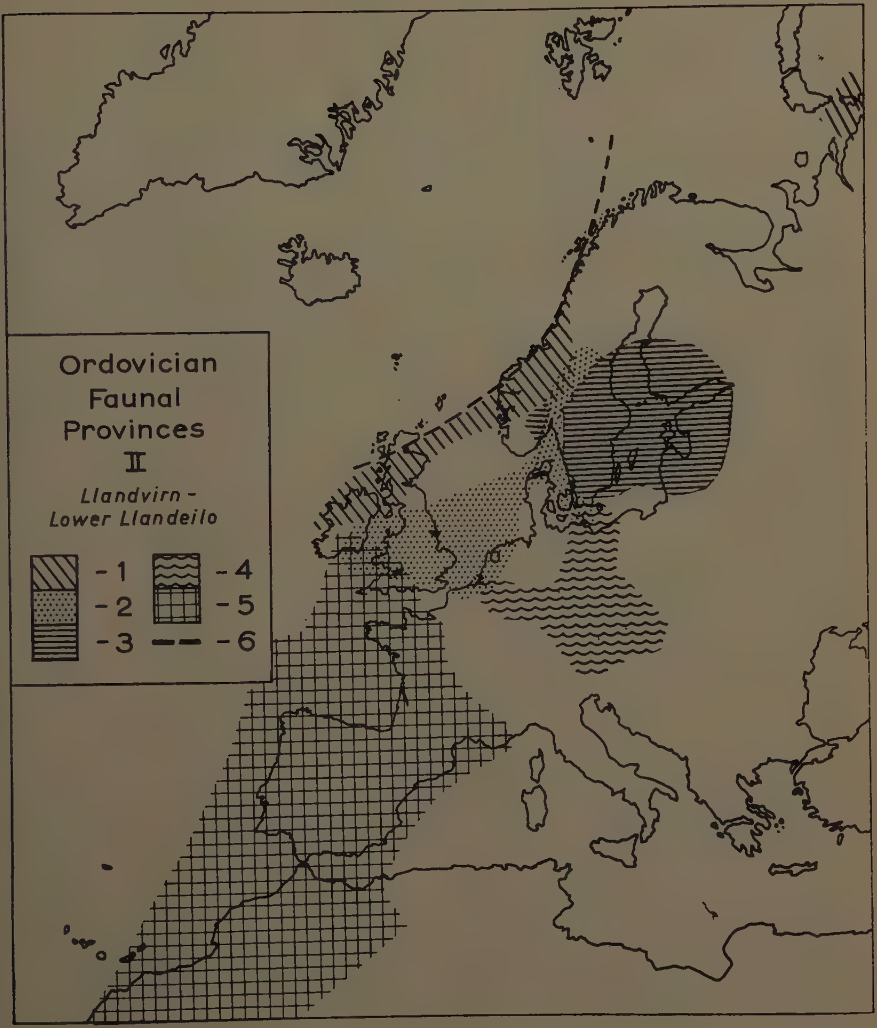


Fig. 2. Ordovician Faunal Provinces in Europe in Llandvirn-Lower Llandeilo. 1 = Faunas of Scoto-Appalachian type, 2—3 = Faunas of Anglo-Scandic type, 2 = mainly graptolite shales and geosynclinal sediments, 3 = shallow-water limestones of Baltic type. 4 = Bohemian faunas, almost undistinguisable from 5. 5 = *Synhomalonotus tristani* province (extends also eastwards to Syria, and westwards to Florida and Alabama, possibly also to Argentina). 6 = approximate limit of American-Arctic province.

Mediterranean province. This province is much more uniform in its faunas than the other ones, and is found in an area reaching from Syria (SUDBURY 1957) in the east to Florida and southern Alabama on the west, and from Belgium, Czechoslovakia and parts of southern Wales on the north, to the Sahara or NW Argentina on the south.

It is typically developed in Bohemia, Normandy, Brittany and Portugal. The "type-fossil" for the province is *Synhomalonotus tristani*. The fauna is composed largely of trilobites, lamellibranchs, and, locally, graptolites. Brachiopods and ostracodes are minor constituents, and cephalopods, bryozoans and corals are very rare. At one horizon, however, in the upper Middle Ordovician, a fauna similar to that in the Anglo-Scandic province invaded the area covered by this province. This fauna contains abundant brachiopods, bryozoans and cystids, some trilobites and a few rugose corals. This is the only limestone horizon in the Mediterranean province.

Except for this bed, all the rocks in this region are dark coloured sandstones, siltstones and shales. In spite of their appearance, the rocks are not generally graywackes; most of them belong to the ortoquartzite suite.

The characteristic phacopid trilobite genus of this province is *Dalmanita*, other genera are *Colpocoryphe*, *Calymenella* and especially *Synhomalonotus*. In the Middle Ordovician, the distribution of the fauna is very uniform, but in the Lower Ordovician it is possible to discriminate environmental (facial) subprovinces. In the western part of Brittany and Portugal, and possibly Western Morocco the Upper Arenig is developed as graptolite shales and siltstones. Then follows a belt of the Armorican Quartzite, a thick light-coloured sandstone, ranging from orthoquartzite to subgraywacke, in some cases with slumping structures. It is found mainly in a N-S belt from Normandy to Morocco. On the eastern side of this belt, the Tremadoc, which like the lower Arenig seems to be missing on the western side, is developed with the *Euloma-Niobe*-fauna (cf. BRØGGER 1896), which is found not only in most parts of Europe, including the Ural Mountains, but also in South America (cf. HARRINGTON & LEANZA 1957). The Arenig is developed in some few regions only(Montaigne Noire and Southern Morocco) as mudstones and sandstones with a trilobite fauna dominated by asaphids.

In the lowest part of the Middle Ordovician, the faunas of this province extended into Belgium, most of Germany, and possibly into Cornwall, and South Wales. In the latter district a mixed fauna is found, this is also the case in S. Eire (Tramore and Raheen Beds). In Bornholm and Scania, some elements of this fauna are also found, especially in the Upper Ordovician.

One of the most interesting occurences of this faunal province is in the subsurface of Florida and southern Alabama. (WHITTINGTON 1953, APPLIN 1951 and BRIDGE & BERDAN 1950). Here, the Ordovician sequence begins with a thick sandstone series (the Armorican Quartzite), which rests upon rhyolite lavas and tuffs, probably corresponding to the Cambrian or Infracambrian volcanics of Brittany. Above the sandstone, there is a series of dark mudstones, which resemble the *Synhomalonotus tristani* beds of Brittany and Normandy lithologically. The trilobite described from these beds by WHITTINGTON (1953) *Colpocoryphe exul*, is a typical Mediterranean one. The graptolites found in the lowest beds of the mudstone series, in the transition to the sandstone, are similar to those found at La Pile-en-Erbray, S. of Rennes in Brittany. One species is common, the age is probably the same, and even the lithology is remarkably similar (cf. PHILLIPOT 1950).

A remarkable fact is that this sequence is not metamorphic, and not folded in spite of its occurring in the continuation of the central part of the Appalachian geosyncline.

Evidence for the Climate in the Ordovician.

The evidence bearing on the climate of the Ordovician is partly of biological, and partly os physico-chemical nature. Since this study is necessarily restricted to marine environments, the problem is to determine the temperature of formation of the shallow-water, shelf deposits.

An absolute determination of the temperature with oxygen isotopes is out of question because all fossils of Ordovician age seem to be too recrystallized to give reliable results.

The known temperature range of living species can be used to determine the climate of fossil deposits in which they are found, but the method is limited to young beds, where living species are still found, and can, therefore, not be used directly in Ordovician beds. It is not possible to use the temperature range of genera and higher taxonomic units, because these are generally too wide, and will give highly inaccurate results. (Might have changed habiat with time!)

There are, however, some generalized rules that might assist in evaluating paleo-temperatures. Large, thick calcareous shells are generally found in warm water. This does not mean that large, thick shells do not occur in cold water; on the contrary, many large, thick-shelled lamellibranchs are found in the arctic fauna. The warm water forms generally, however, have more excess calcification and are, as a rule, more spinose and highly ornamented. This might be connected with the higher solubility of calcium carbonate in cold water, which will benefit forms with the smallest possible surface of the shells. Several cold water shells also have a strongly developed periostracum or live buried in clay, thus preventing corrosion by the sea water.

The quantitative distribution of certain species is also difficult to use when their temperature range is not known. Changes in salinity, clearness of the water and supply of food and oxygen might lead to changes which resemble those of a temperature change. The presence of dwarfed faunas will in most cases indicate lower oxygen content and not lower temperature, whereas a reduction in the number of species and specimens in a fauna is more likely to be the result of changes in salinity or food supply than in temperature. Even if temperature changes might be involved in these ecological variations, they are not reflected directly in the observed facts, and it is very difficult to interpret the role of temperature changes in such cases.

It is still more difficult to determine the temperature of euxinic deposits. They form closed systems with fairly constant temperature, independant of the surroundings, and the fauna consists either of highly specialized forms, or of drifted in planctonic forms. Such environments (the graptolite shales) are therefore not considered here, they might occur in all climatic zones, and are found in the present oceans from the tropics to the arctic.

Within the same species it seem to be a rule that the individuals living in warm water are larger, and have thicker shells than those living in cold water. This is of course valid only if not animals lived in environments of the same type, with equal salinity, oxygen content

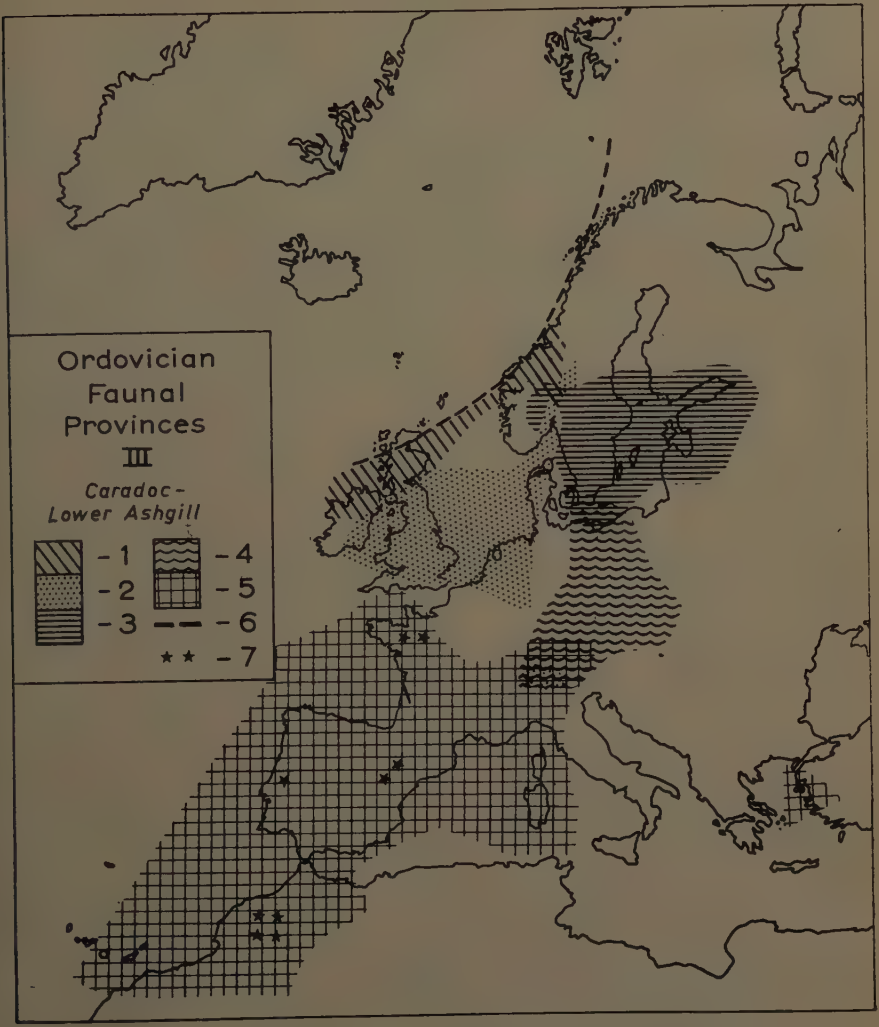


Fig. 3. Ordovician Faunal Provinces in Europe in Upper Caradoc-Lower Ashgill. 1 = Faunas of Scoto-Appalachian type, 2 = Faunas of Anglo-Scandic type (mainly geosynclinal facies), 3 = Baltic type faunas in the Anglo-Scandic province. 4 = Faunas of Bohemian type. 5 = Cystid-Limestone type faunas in the Mediterranean province. 6 = Approximate limit of American-Arctic faunas. (elements of this province are found also in the Anglo-Scandic province in this period) 7 = Occurence of the *Kloucekia*-fauna above, or replacing the Cystid-limestone fauna.

etc. This rule has been used by FLOWER (1946) to suggest that the Red River faunas lived in warmer water than the Cinncinnatian ones. The rule applies, however, only to forms with continuous growth. Forms with growth-stages, moulting, such as most arthopods, are generally smaller in warm water because of the higher rate of growth. This is especially well known from living ostracodes and copepods, and might well have been valid for the trilobites also.

Coral and algal reefs are supposed to be indicators of warm waters and there can be not doubt that the great majority of them were built in warm waters; but exceptions are found, as indicated by TEICHERT (1958). The cold-water coral reefs differ, however, in several respects from the warm-water ones. The sediments concentrated in the "frame-work" of the cold-water reefs are mainly clay and unabraded shells, whereas calcilutite and calcarenite dominate in the warm-water ones, with very little clay. This is because the cold-water reefs are necessarily also deep-water reefs, where the currents do some sorting, but not much abrasion. Because of the depth, the reefs of this type are continuously buried by terrigenous material, whereas the warm-water reefs, being found in shallow water, are seldom reached by the deposition of terrigenous material, and the "framework" sediments are often higly abraded by wave action.

Calcareous algal masses are less distinctly divided into cold and warm water types than the coral reefs, and they are found also at fairly shallow depth in cold water. It is therefore, more difficult to evaluate the temperature of deposition of algal mass deposits, even if the same criteria as for the coral reefs (fragmentation etc.) can be used in some cases, and that the cold water algal masses are more apt to be dissolved, and therefore not preserved. Because of the stronger sunlight, and therefore deeper penetration of light, and the generally clearer water in the lower latitudes, the lime-secreting algae are also more common and more diversified in warm-water.

The annual rate of growth of certain animals, especially corals has also been used to evaluate the relative temperature. (MA 1937, 1956). This method is certainly very valuable, but in order to get reliable results, it is necressary to have a great, and exactly determined material (both stratigraphically and zoologically) in order to eliminate local, ecological variations. This has not yet been done systematically.

The evidence provided by the lithology of the beds will give a more direct temperature than the faunistic evidence, because it might be supposed that the same physico-chemical rules governed the formation of the sediments both in the Ordovician and the present oceans. If, therefore, the process of formation of a type of sediment depends upon temperature, the ancient ones must have been formed at the same temperature as recent ones.

The formation of sediments is, however, governed by other features than temperature, and a purely lithological temperature determination will in most cases be rather uncertain. Generally only a maximum or minimum temperature can be found.

The presence or absence and type of calcium carbonate deposits is probably the best available temperature indicator obtainable for the lithology of the rocks (cf. WISEMAN 1959).

It is well known that calcium carbonate is much more soluble in cold than in warm water, and therefore deposition is not likely to occur in cold water. This does not mean that limestones cannot be formed in cold water, but they will be much less common there than in warm water. In cold water, the calcium carbonate is formed purely biochemically, as shells of organisms. The limestones will consist of coquinas and coarse-grained calcarenites or calcirudites. They are often cemented by clay, but might also have calcite cement, because the pore-water in these sediments easily will have a much higer carbonate content than the sea-water, and therefore permit chemical deposition of calcite. Because the smaller carbonate grains have a proportionally larger surface than the larger ones, they will be dissolved more rapidly, and are seldom found in pure carbonate rocks from cold water deposits. Most fine-grained calcareous rocks have a high clay content, because the clay will prevent the solution of the carbonate, by restricting circulation between the pore-water in the sediments and the sea-water. Corrosion surfaces (diastems) are also more common in cold than in warm waters, but they indicate primarily a drop in temperature, not necessarily to very low ones. They might also be due to invasion of masses of water undersaturated with calcium carbonate, but not much colder than under the previous conditions.

Accretionary calcarenites of the bahamite type definitely indicate warm weater, for they are the result of chemical deposition and contemporaneous abrasion. The fine calcilutites of the drewite type are also indicative of warm water. As mentioned above, calcareous mud deposited in cold water is generally mixed with considerable quantities of clay.

The discrimination of cold and warm water deposits with the criteria mentioned here, is rather difficult, because the thick sediments formed in areas with an ample supply of terrigenous material, will look "colder" than the contemporaneous sediments deposited in areas with the same temperature, but with less deposition of terrigenous material. If it is supposed that the rate of calcium carbonate deposition is constant, the resulting sediments will be much less calcareous in the areas with much terrigenous material, and more mixed with clay than the more purely calcareous deposits in the area without.

This is specially important when comparing subsiding basins (geosynclines) with stable shelt areas. Generally a large amount of observations is needed to get an opinion as a whether the changes are the results of changes in temperature, or if they are due to differences in the supply of terrigenous material In a larger area, this is simpler, just because it is then possible to trace changes parallel to the basin, and compare them with the total thickness of the beds.

The type of clay mineral found in sediments might also give clues to determination of temperature, since different clay minerals are formed by weathering under different climatic conditions. Recently, a number of authors have studied this problem but not general agreement as which clay minerals are indicative of which environment has been reached, except that previously known, that kaolinite and bauxite indicate a warm humid climate, and that cold climates generally are indicated by illite, clorite and muscovite. It seems however, that some clay minerals are changed into others when redeposited in marine environments, and this process is not yet completely investigated. The clay minerals are also primarily formed by terrestrial weathering, and their composition is therefore indicative of the land climate of the erosion area, which is not necessary significant of the temperature of the sea in which the clay minerals were deposited.

Several cases are known in which the temperature of the sea water is much higher or lower than would be expected from the climate of the adjoining land. In most cases this is due to oceanic currents and the topography of the land-masses.

The more intense and rapid weathering in warm water will also affect the feldspar content of sandy or silty sediments. Because the decomposition of the feldspars is mainly a chemical process, they will endure longer transport and more abrasion in cold than in warm water. As far as the author knows no detailed studies of the rate of deposition of the feldspars due to temperature and way of transport has been made, but it seems possible that sorted sandstones and siltstones with a notable amount of clastic feldspars are formed in cold water. The absence of feldspars in unsorted sediments, especially graywackes, is not absolutely conclusive evidence of warm water because the source material might have been poor in feldspars.

It is supposed that graywackes deposited in cold water will show a higher feldspar content than ones deposited in warm water if the source area and the distance and time of transportation is the same. In order to use this feature as a temperature criterion, it is, therefore, necessary to have rather complete data on the geological history and stratigraphy of the region studied.

As mentioned above, sediments in the graywacke facies look "colder" than those in the orthoquartzite facies. Generally it is difficult to discriminate between cold and warm water sediments in the graywacke facies, exept for variations in the feldspar content. Interfingering of calcareous sediments, might in some cases given a clue to the determination of temperature.

In the orthoquartzite facies the cold water rocks will have a superficial resemblance to the graywackes, but a closer examination reveals that the sand- and siltstones are cemented with quartz, and not with clay. The thickness of the beds will in many cases be greater in cold than in warm water, because the chemical weathering and higher solubility of weathering product in warm water will contribute to transport of material in solution, in contrast to the almost purely mechanichal weathering in a cold climate.

It might be concluded that it is not *generally* possible to determine the temperature of formation of a certain sediment from its lithology, but it can be done in a number of special cases. The same conclusion has been arrived at with the biological evidence. If the special cases are numerous enough, it might still be possible to get a fairly reliable basis for the determination of the paleoclimate. In most cases it will not be possible to tell the temperature in a single sample of a single section, a large number of observations, combined with a thorough knowledge of the geological history of the area is necessary.

The best results might be expected where it is possible to use a combinations of lithological and biological evidence, reconstructing types of environment, the temperature of which can be estimated within narrow limits, because they are known also from recent oceans.

Two examples can be mentioned: The presence of rapidly lime-secreting organisms, such as calcareous algae of the family Dasycladaceae in mud-rocks with little or no carbonate in the matrix are indicative of cold water (the algae are used as en example here, instead of the more commonly occuring lamellibranchs, because some of the latter will be burrowing forms).

In warm water the algae-bearing sediments are in most cases highly calcareous, due either to chemical precipitation, or to fragmentation of biologically precipitated carbonate. Only in cold water there will be both an ample supply of dissolved carbonates, which can be utilized by the plants and still no chemical precipitation, and no deposition of clastic carbonate material. In the few and special cases in warm water where the supply of clay is large enough to suppress the carbonate precipitation completely (in deltas of large rivers) there will generally be features indicating the special conditions, such as impoverished or specialized faunas due to low or variable salinity and possibly deposition of silica and iron, which are less soluble in salt than in fresh water.

The other example is the pure, almost unfossiliferous calcilutites ("dove limestones", "birdseye limestones") in which the fauna is restricted mainly to smooth ostracodes and gastropods. They are in most cases interbedded with thin layers of very fossiliferous calcarenites. This type of environment correspond perfectly to the calcareous muds known from shallow water shelf and lagoons in warm water, i.a. from island in the Pacific, and especially from the Bahamas. This environment is characterized by chemical deposition of the calcerous mud (drewite), and by its very poor fauna, due to lack of nutrients. The calcarenitic beds are due to occasional periods with currents, resulting in erosion of the drewite, and supporting a rich bottom life. A typical feature of these calcarenite beds is that the lower surface is a sharp, erosional contact, and the upper one is gradational into the calcilutite.

"Calcilutites" from somewhat colder water, are seldom purely lutitic, they are mixed with calcarenites, often in small lenses and nests, which might have a considerable vertical extent. If they are deposited in colder water, they often have a considerable content of clastic clay.

The Position of the Poles during the Ordovician.

The limitation of the climatic zones (=faunal provinces) is not always quite clear, and might be subject to discussion, but the trend in their arrangement, from warm to cold is obvious. In Europe this is from N and NW to S, and in N. America it is from NW and W to E and S. This clearly indicates that the location of the poles and the equator was quite different in the Ordovician from their position at present. The fact that the so called "Arctic" Ordovician faunas were warm water ones has been pointed out by a large number of authors, the first was possibly HOLTEDAHL (1920), and strong faunal evidence has been given i.a. by FLOWER (1957 a.o.). The logical interpretation of this is that the equator passed over the present arctic, and the poles must therefore have shifted almost 90°.

The question of the stability of the poles, and of their movement has been thoroughly discussed, as though no general agreement has been reached. Some other evidence for shift of the poles is obtained from study of the Cretaceous climatic zones, which seem also to have been inclined obliquely to the present ones, and in the same direction as the Ordovician ones.

Salt deposits are generally supposed to form in the trade wind belts (15—25° N or S), at least on the northern hemisphere (LOTZE 1938 pp. 154—159, textfig. 78). The distribution of the salt deposits in the Mesozoic and Paleozoic seems to indicate a shift of the poles similar to that suggested here, even if the amount of movement is supposed to be less than 90° (LOTZE l.c.).

Some data regarding polar shifting come from paleomagnetic studies (cf. RUNCORN 1959).

Paleomagnetic research is in a very rapid development, and the constant reevaluation of important principles makes the uncritical use of all published data somewhat dangerous. There is, however a general agreement on the point that it is possible to determine the

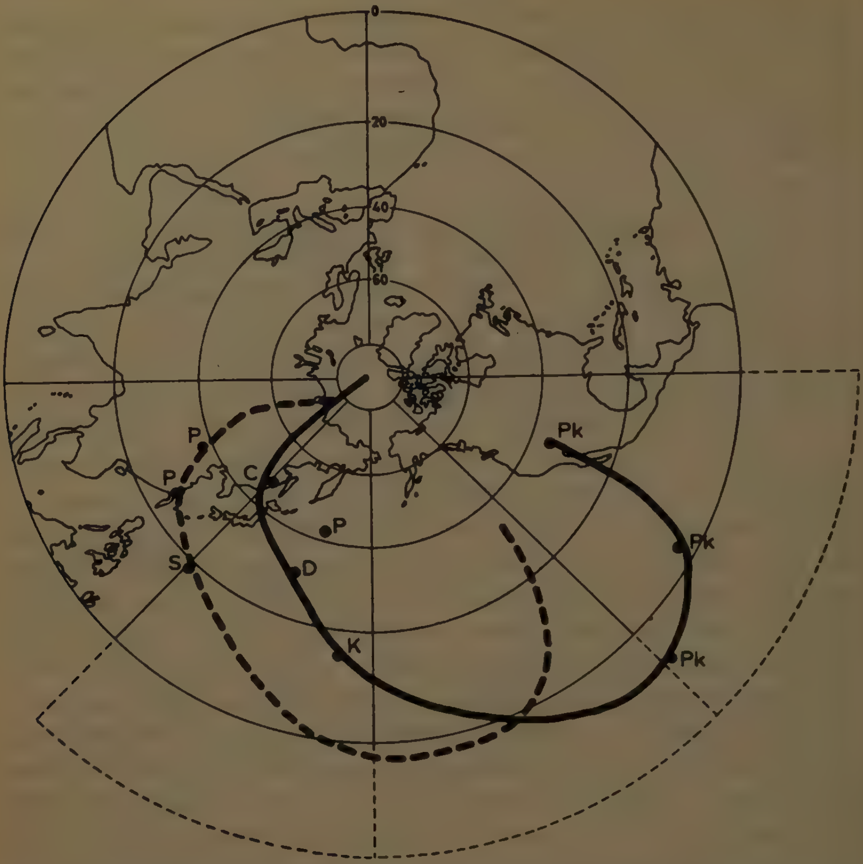


Fig. 4. The position of the magnetic north pole according to paleomagnetic data. Unbroken line indicate the path of the magnetic north pole based on observations in Europe, the broken line the path for North America. P = Permian, C = Carboniferous, D = Devonian, S = Silurian, K = Cambrian, Pk = Precambrian. Simplified from Runcorn 1959.

position of the magnetic poles of an earlier period in geologic history by studying the remnant magnetism of rocks from that period. It is also generally accepted that the poles have moved during geologic history, and most published data seem to indicate that the general movement of the magnetic north pole can be traced back, accross the Arctic Ocean, and southward to Korea or North China, where possibly it was located in the Permian and upper Carboniferous (RUNCORN 1959, cf. textfig. 4).

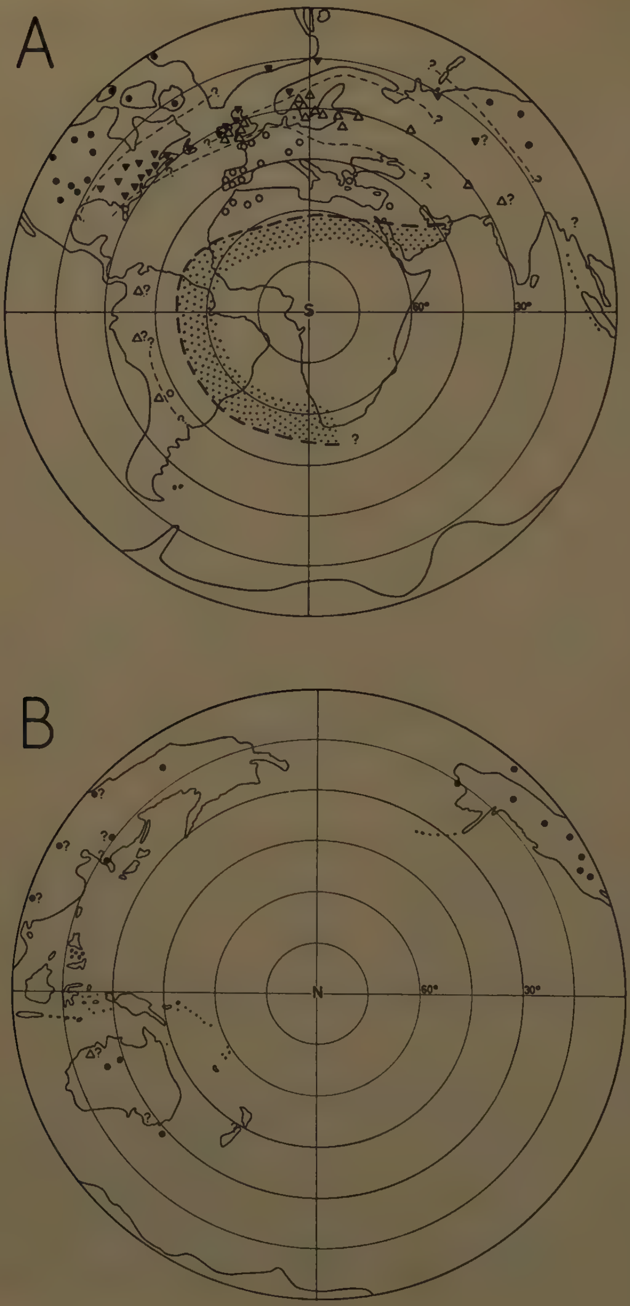
The data for older periods are very scanty, and not consistent in details. It seems however to be a trend in the same direction as in the later periods. An extrapolation would give the magnetic north pole not far from equator, possibly somewhat to the north of it, and at about 165° E. This estimate correlates well with the few, and admittedly inexact data for the Lower Paleozoic. The figures mentioned here are highly inaccurate, but it seems probable that the Ordovician magnetic north pole was from 0 to 25° N. of the present equator, and at between 150° E and 180° W.

The paleomagnetic determination of the magnetic poles is not a determination of the axis of rotation of the earth. At present the magnetic north pole is ab. 15° from the geographic pole, and there is no reason to believe that this distance was smaller in the Ordovician. It is, probable, however that the geographical pole in the Ordovician was also near equator, somewhere in the central Pacific.

The antipode of this pole would have been somewhere in, or W. of Africa, which corresponds surprisingly well with the picture obtained from the distribution of climatic zones, with warm water in the N. and NW. in Europe, being gradually colder towards S., and similar conditions in N. America. It might therefore be concluded that the paleomagnetic evidence does at least not contradict the arrangement of climatic zones suggested here.

The question of whether the movements of the poles indicate a movement of the whole mass of the earth, or whether such movement means that only the continents have moved in relation to the axis of rotation, is a problem which is outside the scope of this paper.

Continental drift also has to be mentioned in connection with the arrangement of the climatic zones. The climatic zones in Europe and North America do not fit well together with the present position of the continents, and a much better agreement is obtained if they are moved closer together. In the maps (textfigs. 5A—B) the two continental blocks on both sides of the Atlantic are moved about 40° closer, and this gives the best possible arrangment with the data available. This must not be regarded as any proof of continental drift, it is purely an empirical geometrical adjustment. There are too many variables in the reconstructions behind these maps to regard them as conclusive


A
S
60°
30°
Δ?
Δ?
?
B
N
60°
30°
Δ?
?

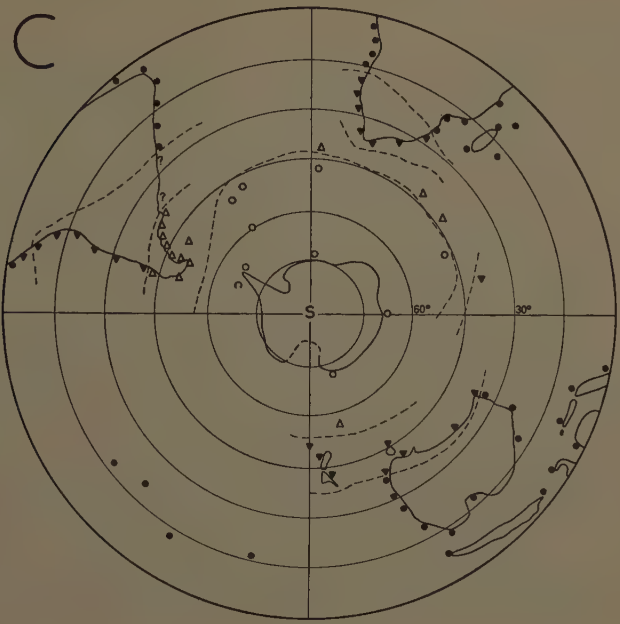


Fig. 5. A-B: Reconstruction of the climatic zones and geography of the Ordovician (Mainly Middle Ordovician) based on the distribution of the faunal provinces. C. The recent faunal provinces on the southern hemisphere, introduced for comparison. (Mainly after EKMAN 1953).
Filled circles = Tropical faunas, Filled triangles = Warm temperate faunas, Open triangles = Cold temperate faunas, Open circles = Antarctic and Arctic faunas.

evidence for any theory, exept that of climatic zoning. The pole might have been farther to the west, and the climatic zones are not strictly parallel to the latitudes, oceanic currents are responsible for considerable discrepancies (cf. textfig. 5 C). Such discrepancies were certainly present in the Ordovician also. If all these sources of error are biased in the same direction, the resulting inaccuracy might reach the same order of size as the supposed continental drift. The amount of continental drift, if present at all, can therefore not be determined from the data available.

It has also been suggested that India and Australia have drifted apart from Africa, but no evidence is found for that in the study of the climatic zoning. The Ordovician in India is along the edge of the old continental block, in the Himalaya and Burma, and it is not well enough known to give much information on the climate. However, it possibly belongs to neither the very cold nor the very warm areas.

Australia might have drifted from a position symmetrically on the other side of the Ordovician equator, that is about 60° nearer Africa, but no direct evidence is found for this drifting either.

It might be concluded that the arrangement of the climatic zones in the Ordovician gives no direct evidence for or against continental drift. It is possible that a more detailed study of the distribution of the faunal zones might give some evidence, especially the striking lithological resemblance between the rocks in Brittany-Normandy and in the subsurface of Florida and Alabama. Many studies have to be made, however, before any conclusions can be reached. The only positive thing that can be said about continental drift based on the evidence present in this paper is, that if continental drift has occurred since the Ordovician, it must have been in the direction generally supposed, away from Africa, and generally in an E-W direction. Considerable movements of the continents in other directions would not fit with the picture of the climatic zoning obtained here.

The number of common species and genera in the Ordovician on both sides of the Atlantic might give some evidence on the possible amount of continental drifting. In the recent Atlantic a sharp zoogeographical boundary is found at Cape Cod, at the NE-ern coast of U.S.A. North of this point, the arctic fauna is found with a higher percentage of genera and species which are also known on the other side of the Atlantic. South of Cape Cod there are comparatively few common genera and species in the boreal fauna. This is probably due to the short or almost nonexistant distance between the shelfs in the Arctic Basin and its surroundings, and the long distance between the shelfs across the Atlantic at the latitude of Cape Cod. Even if the situation is complicated by the system of oceanic currents, it might be interesting to compare the percentages. According to Ekman (1953 p. 140) the percentage of common species in two animal groups (fishes and ascidians) are 50% and 88% N. of Cape Cod, and 8% and 9% S. of it. When these figures are compared with the number of common genera and species between the Appalacians and Scotland and Scandinavia in the Ordovician, it is evident that the figures at least in the best known groups of benthonic animals (trilobites and brachiopods) are much nearer to those of the arctic than to the boreal faunas. This seems to indicate that the distance between the continental shelves of Europe and North America was considerably smaller in the Ordovician than it is now. It must be mentioned, however, that it might not be justified to compare the precentages between the recent and fossil genera and species, because they are probably not biologically equivalent. Further, and more detailed studies along these lines might give more exact data to evaluate the presence and amount of continental drift.

An interesting feature seen from the maps (textfig. 5A—C), is that the Ordovician climatic zones do not seem too much distrupted by later orogenies. In Europe, the climatic zones are cut at low angle by three mountainchains, the Caledonian, the Hercynian, and the Alpine, and the total compression seems to be smaller than the displacements which can be ascribed to oceanic currents.

The figure given for the shortening of the upper crust by the formation of mountain chains are highly variable. The highest for the Alpine folding are several thousand kilometres. Judging from the position of the Ordovician climatic zones, these figures must be vastly exaggerated. The right figure is probably a few hundreds of kilometeres.

Variation in Climate during the Ordovician.

The climate was probably not stable through the whole Ordovician, climatic changes were numerous and important. In fact, the exact limitation of the climatic zones is made difficult by their change with time.

If the temperature increases, the warm water faunas will migrate towards the poles, and a parallel migration will be found in the intermediate faunas. When the climate becomes colder, the reverse migrations will take place, the cold, and intermediate faunas will migrate towards equator, (cf. textfig. 6).

Judging from the migrations of the N. European marine lamellibranch fauna in post-glacial times, the movements of the faunas are wide, and take place very fast, geologically speaking. The *Tapes*-fauna, indicating the post-glacial climatic optimum in Scandinavia, migrated more than 1000 km towards the north pole, and back in less than 3000 years. It is suggested that most observed faunal migrations in the Ordovician are due to such climatic changes.

Since these changes are contemporaneous over at least one hemisphere, they are of high potential value for long distance correlations. At present, it is difficult to identify the specific climatic changes in all areas, due to lack of detailed studies.

The sharp climatic zoning found both in the Ordovician and at present, is possibly due to the presence of polar ice-caps. That has been indicated i.a. by EWING & DONN (1956, 1958). If no polar ice-caps were present, the oceanic current system would have been considerably different, large part of the oceans might have been stagnant, and the earth as a whole would have a much more uniform climate, with no sharp zoning. Such periods are possibly the Devonian (including parts of the Silurian), the Jurassic, and the Cretaceous. These are all periods with a rather warm, and uniform climate.

The periods with sharp climatic zoning coincides generally with the periods of glaciation, such as the Eocambrian, the Permo-Carboniferous and the Pleistocene-Recent. These periods of glaciation are also accompanied by maximal orogenic activity, in the two last cases mentioned by the Hercynic and Alpine orogenic system. The Caledonian orogenic system which had its most active phases in the Ordovician is exceptional, in being not accompanied by a regional glaciation. This might partly be due to the long duration of that orogenic system, from the base of the Ordovician to the Middle or Upper Devonian. The Ordovician climate was probably just as cold as in the other periods, in which glaciations are found, but the meteorological conditions were probably different, and high mountains near the ocean coasts, necessary for the accumulation of a continental ice-

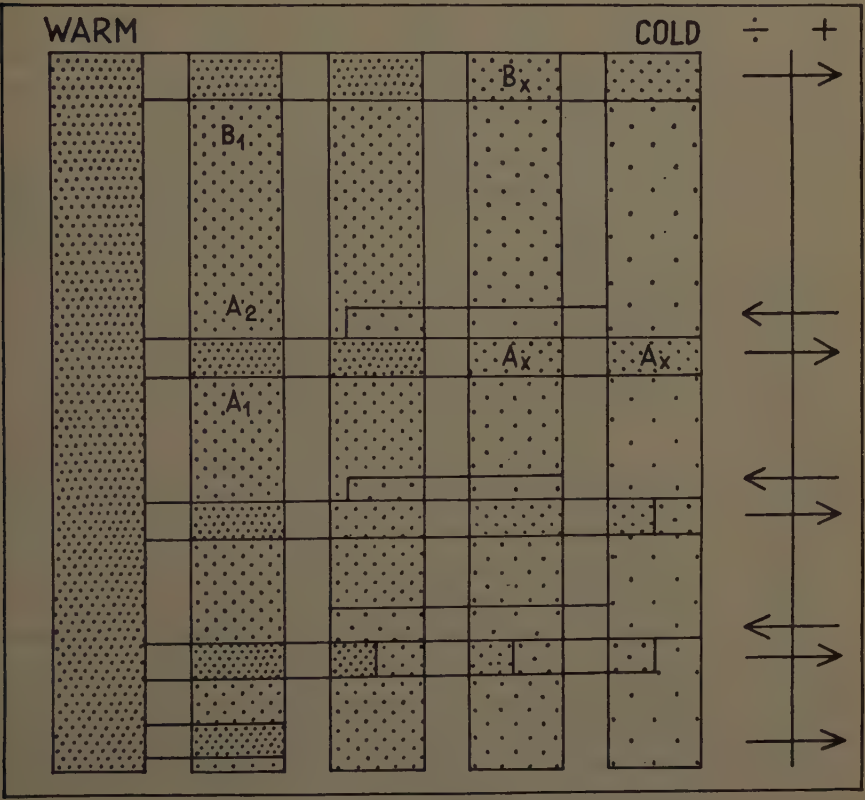


Fig. 6. Diagram showing the faunal sequence in a number of hypotetical sections due to faunal migrations induced by climatic changes. Denser punctation indicate warmer faunas. The migration of warm faunas into "cold" areas and vice versa due to change in climate is illustrated. Faunas like Ax are not strictly contemporaneous with the highly similar A^1 or A^2, they should be correlated with the "warm" fauna between the two latter. In this way the climatic changes might be used for long distance correlations.

cover, were probably not present. This will explain both the absence of a glaciation in connection with the Caledonian foldings, and the presence of sharp climatic zoning in the Ordovician.

The presence of polar ice-caps has a great stratigraphic importance. The climatic changes will increase and decrease the volume of water bound in the ice-caps, and that would lead to transgressions in warm periods, when the ice-caps are reduced, and to regressions in cold periods, when more water is bound in the ice-caps. This climatic dependence of the transgressions and the regressions is especially well illustrated by the changes in sea level known in connection with the last glaciation. There are certainly other causes for changes in sea level, but with the widespread occurence of shallow continental shelves known from the Ordovician, the climatically induced ones must have been very important.

As a conclusion, the author will attempt to give a brief outline of the climatic development during the Ordovician (diagramatic in textfig. 7).

The climatic conditions in the lower and middle Lower Ordovician are not well known, but seem to be warm to intermediate. The upper part of the Lower Ordovician was very warm, the "warm-water" faunas had a wide distribution (Canadian in N. America, *Orthoceras* Limestone in N. Europe). The beds are also transgressive in several areas, such as S. Norway, Portugal and N. Africa.

The lower part of the Middle Ordovician is marked by cold water, a very abrupt change from the preceding stage. In Europe the beds are regressive only locally, but in N. America, the largest Ordovician regressions is found at this time, and the beds, (Whiterock Stage) which are found only along the margins of the continental mass, in Newfoundland, Oklahoma and parts of Utah and Nevada, indicate somewhat colder conditions than in the beds below.

In the Llandeilo, and the Caradoc, the temperature seem to have increased slowly and regularly, except for some fluctuations about the time of the *Climacograptus peltifer*-zone (a slight maximum, and later a slight minimum). An important temperature maximum is found in the upper part of the Middle Ordovician, in the zone of *Dicranograptus clingani*. The beds are locally transgressive, especially in S. Europe and in N. Europe there are a number of limestones with an american fauna (mainly Appalachian, but also some "Arctic" faunas). The Vasalemma in Esthonia, Kullsberg in Sweden, Encrinite, Mjøsa and Kalstad in Norway, Craighead in Scotland, Keisly in England are the typical representatives. In S. Europe there was an invasion of the fauna which previously lived in N. Europe, dominated by bryozoans, cystids and brachiopods. At this level the only limestone horizon occur in this otherwise purely cold-water province.

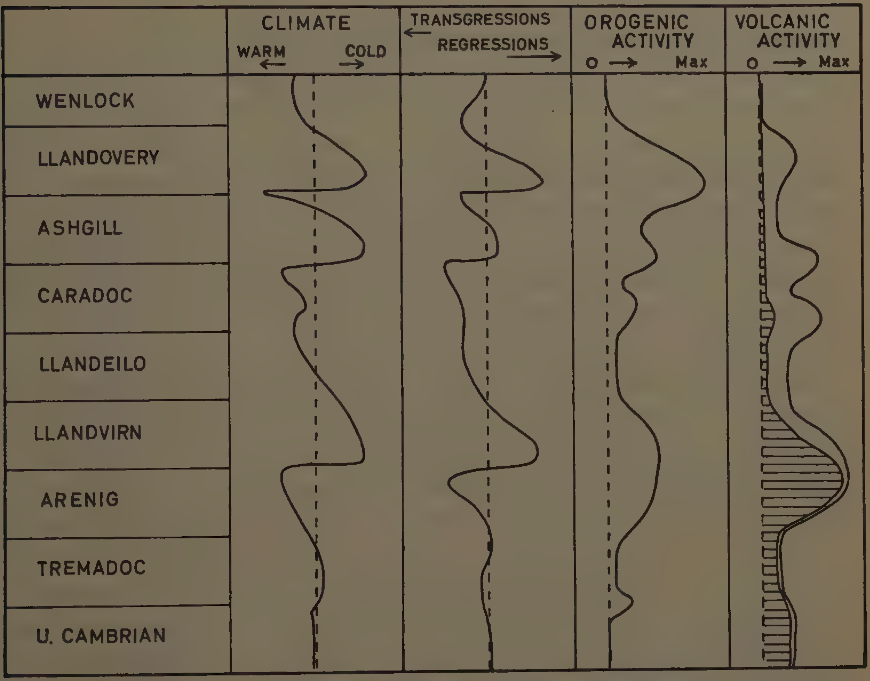


Fig. 7. Diagram illustrating the correlations between climate, transgressions and regressions, orogenic activity and volcanic activity in the Ordovician and parts of the Cambrian and Silurian.
The data for climate, transgression and regressions from the whole world, those for orogenic activity and volcanic activity are mainly from Europe.The shaded part of the column for volcanic activity indicate approximately the fraction of basaltic volcanic activity. The strength of the orogenic activity is indicated by the degree and distribution of metamorphism, and the distribution of distinct angular disconformities. The strength of the volcanic activity is taken as the approximate volume of extruded lavas and tuffs.

The maximum is abruptly followed by a minimum in the zone of *Pleurograptus linaris*. Return of "colder" faunas, and widespread regressions, especially in Europe, and in Eastern and Western N. America marks this stage.

During the Ashgillian, the temperature seems to have risen gradually, until it reached a new maximum in the uppermost Ordovician. In N. Europe this stage is defined by limestones with coral-reefs, and faunal elements from the "Arctic" (= Red River) province, especially among the cephalopods. The Porkuni of Esthonia, the Boda and Slätdal of Sweden, the 5b ot Norway are typical examples. In North America the climatic changes are less sharp than in Europe, but it is possible that the recurrent faunas represent climatic changes, so that the comparatively cold Maysvillian represent the temperature minimum in the low Upper Ordovician, and the warmer Eden and Richmond represent the maxima. This has already been suggested by ULRICH (1910), who refered to these faunas as "austral" and "boreal". (cf. SWEET, TURCO, WARNER & WILKIE 1959).

The maximum in the uppermost Ordovician is followed by a sharp minimum with widespread regressions in the basal Silurian. Then came a gradual increase through the Middle and Upper Llandovery, accompanied by a great, worldwide transgression.

A distinct, but crude correlation is also found between the orogenic periods, periods of volcanism and the climatic changes. The first major orogenic period in the Ordovician ranges through Arenig and Llandvirn, and is accompanied by strong basaltic volcanism in the Arenig, and more local, mainly andestitic volcanism in the Llandvirn. This corresponds well with the climatic changes at the Arenig-Llandvirn border. A slight disturbance is found in the lower Caradoc, accompanied by considerable andesitic volcanism. The next major orogenic maximum is found in the Upper Caradoc and lower Ashgill. (Ekne phase of VOGT (1945), probably the same as the Sardic and Pallaresic ones, and movements in eastern N. America). This orogenic maximum is accompanied by considerable, mainly rhyolitic volcanism, and correspond to the climatic change between the *Dicranograptus clingani* and *Pleurograptus linearis* zones. The border between the Ordovician and the Silurian is marked by an orogenic maximum, ranging from the Middle Ashgill to the Upper Llandovery. It corresponds to the climatic change on the Ordovician-Silurian boundary.

The following empirical conclusions might be drawn from the data presented above:

1) The increase in temperature seems to be a slow process, whereas the decreases are much more rapid.

2) A definite correlation is found between the major transgression and regressions, and the climatic changes. The regressions seem to

follow immediately after the decrease in temperature, but the transgressions come sometime in the long interval of temperature increase.

3) A distinct correlation is found between the orogenic maxima and the temperature changes. The maximal temperature is generally reached after the orogenic period had started, and the orogeny continues after the temperature change. The climatic changes are therefore supposed to be better time-indicators than the orogenies.

4) The correlation between the volcanic activity and the climatic changes is very crude, and possibly a secondary feature, due to the correlation of volcanism with the orogenies. There is no quantitative correlation between the volcanic activity and the climatic changes, and the maximum of volcanism during an orogenic maximum might come before, after, or be contemporaneous with the climatic change.

As mentioned above, these conclusions are purely empirical, and can at present only be used for the Ordovician, because they are based on a small number (3—4) of correlated cycles.

LIST OF REFERENCES

APPLIN, P. L. 1951,: Preliminary Report on buried Pre-Mesozoic Rocks in Florida and adjacent States. U.S. Geol. Survey, Circular *91*, pp. 1—28. Washington D.C.

BRIDGE, J. & BERDAN, J. M. 1950, Preliminary Correlation of the Paleozoic Rocks from test wells in Florida and adjacent parts of Georgia and Alabama. Unpublished Report, U.S. Geol. Survey. pp. 1—8, 1 map. Washington D.C.

BRØGGER, W. C. 1896: Über die Verbreitung der Euloma-Niobe Fauna (der Ceratopygenkalkfauna) in Europa. Nyt Mag. Nat.-Vidensk. *35*, pp. 164—240 Kristiania (Oslo).

COOPER, G. A. & KINDLE, C. H. 1936, New Brachiopods and Trilobites from the Upper Ordovician of Percé, Quebec. Journ. Paleont. *10*, pp. 348—372, pls. 51—53. Menasha, Wis.

EKMAN, S. 1953, Zoogeography of the Sea. pp. 1—417. Sidgwick and Jackson Ltd. London.

EWING, M. & DONN, W. L. 1956, 1958, A Theory of Ice Ages I—II. Science *123*, pp. 1061—1066 and *127*, pp. 1159—1162. Washington D. C.

FLOWER, R. H. 1946, Ordovician Cephalopods of the Cincinnati Region, Part I. Bull. Amer. Paleont. *116*, pp. 1—656, 50 pls. Ithaca, New York.

— 1957, Studies of the Actinoceratida. Mem. New Mexico Bur. Min. Res. *2*, 102 pp. 13 pls. Socorro, N. Mex.

FOERSTE, A. F. 1921, Notes on Arctic Ordovician and Silurian Cephalopods chiefly from Boothia Felix — King William Land, Bache Peninsula and Bear Harbour. Journ. Denison Univ. Sci. Lab. *19*, pp. 247—306, pls. *27*—35. Granville, Ohio.

HARRINGTON, H. J. *in* JENKS, W. F. 1956: Geology of South America. Mem. Geol. Soc. America *65*. New York.

HARRINGTON, H. J. & LEANZA, A. F. 1957, Ordovician Trilobites of Argentina. Spec. Publ. Dept. Geol. Univ. Kansas. *1*, 276 pp. Lawrence, Kansas.

HINTZE, L. F. 1952, Lower Ordovician Trilobites from Western Utah and Eastern Nevada. Bull. Utah Geol. Min. Survey *48*, 249 pp., 28 pls. Salt Lake City.

HOLTEDAHL, O. 1919, On the Paleozoic Series of Bear Island, especially on the Hecklahook System. Norsk Geol. Tidsskr. *5*, pp. 121—148, pls. 12—15. Kristiania (Oslo).

— 1920, Paleogeography and Diastrophisms in the Atlantic-Arctic Region during Paleozoic Time. Amer. Journ. Sci. 4th ser. *49*, pp. 1—25. New Haven, Conn.

LOTZE, F. 1938, Steinsalz und Kalisalze, Geologie, *in:* STUTZER, O.: Die Wichtigsten Lagerstätten der "Nicht-Erze". b. *3*, teil 1. Geb. Borntraeger, Berlin.

MA, TING YING H. 1937, On the Growth Rate of Reef Corals and Its relation to Sea Temperature. Paleont. Sinica, ser. B. *16*, fasc. 1. 326 pp, 100 pls. Peking.

— 1956, Climate and the Relative Positions of Continents during the Ordovician. Research on the Past Climate and Continental Drift *11*, 29 pp. 11 pls. Taipei.

PHILLIPOT, A. 1950, Les graptolites du Massif Armoricain. Etude Stratigraphique et Paleontologique. Mem. Soc. Geol. Min. Bretagne *8*, 295 pp. Rennes.

POULSEN, C. 1937, On the Lower Ordovician faunas of East Greenland. Medd. om Grønland *119*, no. 3. 72 pp. 8 pls. København.

— 1951, The Position of the East Greenland Cambro-Ordovician in the Paleogeography of the North-Atlantic Region. Medd. Dansk Geol. Foren. *12* pp. 161—162. København.

ROSS, R. J. jr. 1951, Stratigraphy of the Garden City Formation in Northeastern Utah, and Its Trilobite Faunas. Bull. Peabody Mus. Nat. Hist. *6*, 161 pp. 36 pls. New Haven, Conn.

RUKAVIŠNIKOVA, T. B. 1956, Brachipodi ordovika Cu-Ilijskich gor. *In:* Ordovik Kazachstana II, Trudi Geol. Inst. Akad. Nauk SSSR, pp. 105—168, 5 pls. Moskva.

RUNCORN, S. K. 1959, Rock Magnetism. Science *129*, pp. 1002—1012. Washington. D. C.

SCHWARTZBACH, M. 1950, Das Klima der Vorzeit, 211 pp. Stuttgart.

SPJELDNÆS, N. 1957, The Middle Ordovician of the Oslo Region, Norway. *8*. Brachiopods of the Suborder Strophomenida. Norsk Geol. Tidsskr. *37*, pp. 1—214, 14 pls. Bergen.

— 1958, Ordovician Climatic Zones. Program *1958* Ann. Meeting Geol. Soc. America. p. 143 (abstract). New York.

STØRMER, L. 1953: The Middle Ordovician of the Oslo Region, Norway, I, Introduction to Stratigraphy. Norsk Geol. Tidsskr. *31*, pp. 37—141, 6 pls. Bergen.

SUDBURY, M. 1957. *Diplograptus spinulosus* sp. nov., from the Ordovician of Syria. Geol. Mag. *94*, pp. 503—506. Hertford.

SWEET, W. C., TURCO, C. A., WARNER, E. JR., and WILKIE, L. C. 1959. The American Upper Ordovician Standard. I. Eden Conodonts from the Cincinnati Region of Ohio and Kentucky. Journ. Paleont. *33*, pp. 1029—1068, pls. 130—133. Menasha, Wis.

TEICHERT, C. 1958. Cold- and Deep-Water Coral Banks. Bull. Amer. Ass. Petrol. Geol. *42*, pp. 1064—1082. Tulsa, Okla.

TRUMPY, D. 1943. Pre-Cretaceous of Columbia. Bull. Geol. Soc. America *54*, pp. 1281—1304. New York.

ULRICH, E. O. 1910, Revision of the Paleozoic Systems. Bull. Geol. Soc. America *22*, pp. 281—680, pls. 25—29. New York.

VOGT, T. 1945, The geology of part of the Hølonda-Horg District, a type Area in the Trondheim Region. Norsk Geol. Tidsskr. *25*, pp. 449—528, 9 pls. Oslo.

WHITTINGTON, H. B. 1953. A new Ordovician trilobite from Florida. Breviora *17*, pp. 1—6, pl. 1, Cambridge, Mass.

WISEMAN, J. D. H. 1959, The relations between paleotemperature and Carbonate in an equatorial Atlantic pilot core. Journ. Geol. *67* pp. 685—690, Chicago.

Manuscript received June 7, 1960
Printed March 1961

NOTISER

A new silicified coral from the Upper Ordovician of the Oslo Region,

By

NILS SPJELDNÆS

Rugose corals with polygonal (mainly triangular) and elliptical cross section have recently been described from several areas, and their taxonomy and phylogeny has been discussed (DUNCAN 1957, OLIVER 1958). In the upper Ordovician they are good guide fossils (DUNCAN 1956). The presence af a calceoloid silicified rugose coral from zone 5b in the Oslo Region might therefore be worth notice.

Silicified fossils are rare in the Cambrosilurian of the Oslo Region, the only previously known occurence is in zone 5b at Ullerntangen in Ringerike where compound corals are commonly silicified in certain beds. The silicification of rugose corals faciliate the study of the external morphology and the development of the septae, and the generally abundant material can make a statistical study possible. The internal structure especially of the septae are, however, destroyed by the same process.

The present material come from zone 5b in a road section on road 334 just at the junction with the road to the Hærøya factories, at Gunneklev, some 3 km S of Porsgrunn in the Skien-Langesund District, the southernmost part of the Oslo Region. The silicified fossils are found in a bed about 30 cm thick, which is found just below a coquinal bed full of *Holorynchus giganteus*, the pentamerid brachiopod which is the guide fossil for the zone. Some other coquinal beds of this type are found higher upp in the section. The youngest beds of 5b which are exposed are coral bioherms and conglomerates, ab. 7 m above the silicified corals. Silicified fossils might possibly be found also at other levels and localities. No systematic search for them has been made.

The majority of the silicified fossils are rugose corals, and among them more than 95% belonged to the new species, *Holophragma duncanae*. Among the other corals were an unidentified heliolithid

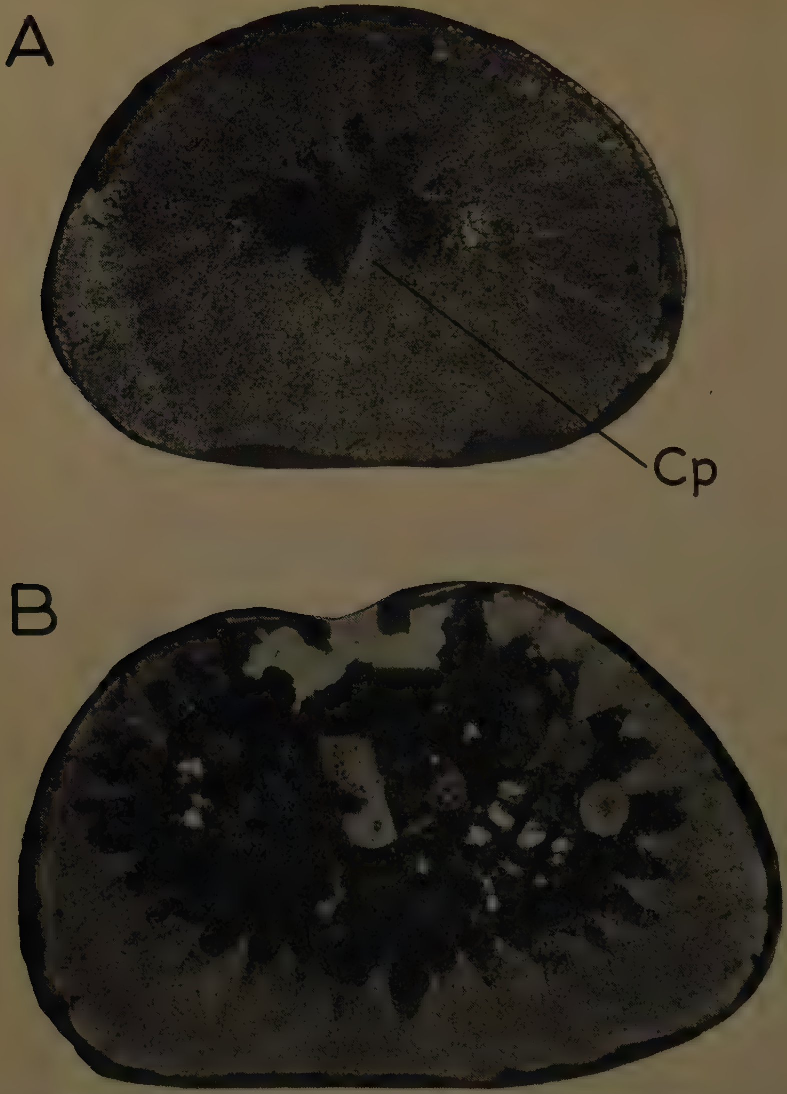


Fig. 1. *Holophragma calceoloides* LINDSTRÖM, from the Visby Marl (Upper Llandovery), at Ireviken, Gotland, Sweden. A. is a deep section showing the columella formed by swelling of the central part of the cardinal septum. The columella less developed, and lower than in *H. duncanae* sp. n. Cp = Columella. B. is a high section, corresponding to that figured by LINDSTRÖM (1896, pl. 6 fig. 85). The section passes above the columella. The light spots in the centre of the picture are fossil fragments. Both sections are photographed from dry peels of the same specimens (PMO A 28139), and are ab. 10 x.

species forming thin crusts, and some few fragments of a *Paleofavosites* sp. The other silicified fossils, which are few, consist of fragments of *Holorynchus giganteus*, which is not completely silicified, whorls of gastropods, fragments of a strophomenid brachiopod, and one specimen of a rhomboporoid bryozoan, and a hypostome, possibly belonging to a Lichid trilobite.

Genus *Holophragma* LINDSTRÖM 1896.

Not much has to be added to the description of this genus given by LINDSTRÖM (1896, pp. 35—37, figs. 74—86) and DUNCAN (1957, p. 609). The only observation which has to be added to the diagnosis of this genus is that a columella consisting of the swollen central part of the cardinal septum is regularly present. LINDSTRÖM's statement (1896, p. 36) that the genus has no columella, might due to a misinterpretation of the figured thin section (l.c. pl. 6 fig. 85). In this section the columella is not shown, because the section do not cut through the central part of the coral. A deeper section from a topotype (fig. 1) clearly show the columella.

Holophragma duncanae sp. n.

(textfigs. 2—3).

Diagnosis: Holophragma species with thin septa and columellar structure consisting of a blade-like expansion of the cardinal or counter septum and variable amounts of vesicular tissue.

Type data: The holotype, PMO 72152 is an almost complete, silicified specimen from zone 5b, at Gunneklev, Skien-Langesund district, Norway.

Material: About 80 complete specimens, and a large number of fragmentary ones, all silicified.

Description: The apical part of the corallae are irregularly cylindrical to narrowly conical. At a diameter of ab. 1,5 mm they expand rapidly, getting a calceoloid shape. Later again, in the mature or gerontic stage of growth, they expand more slowly, and attain an almost cylindrical shape. The development of the shape possibly indicate that the mode of living changed from erect to resting on one side, finally with the calyx turned from the bottom.

The external ornamentation is not well preserved, and only rather coarse longitudinal striation is observed, especially in the apical part of the corallae. The strong external ridge marking the cardinal septum in *Holophragma calceoloides* (cf. LINDSTRÖM 1896, pl. 6 fig. 77) is not found in this species. Growth-lines are not prominent.

The number of septae is 39 in the holotype, in more mature specimens the number may reach 47. The septae form thin discrete plates, the cardinal one is often slightly more prominent than the others.

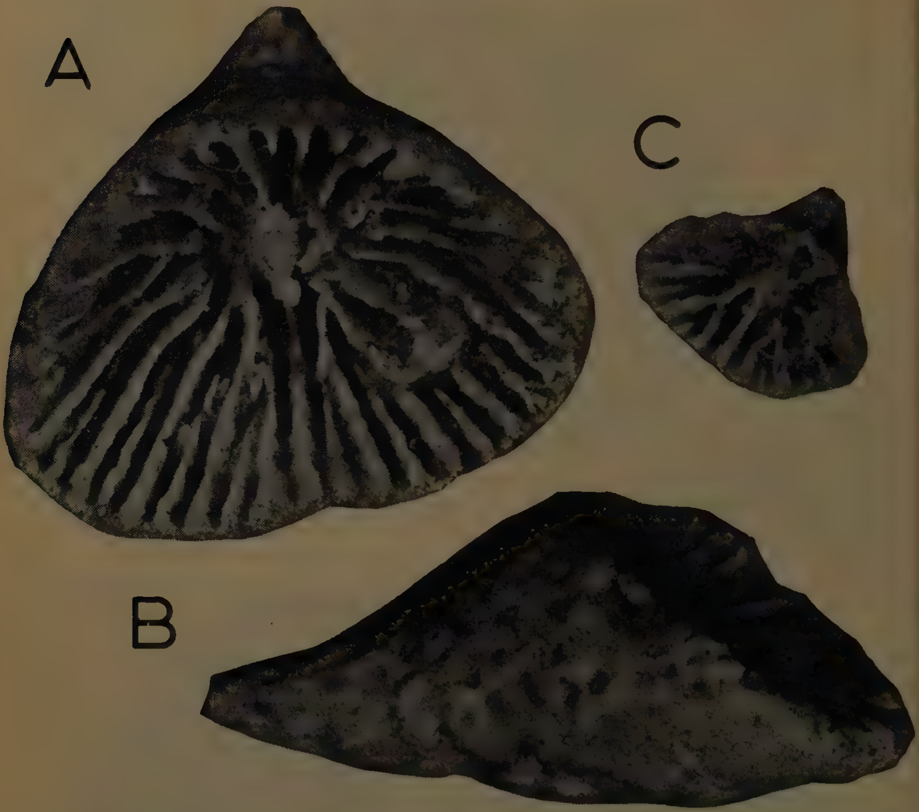


Fig. 2. *Holophragma duncanae* sp. n. from zone 5b (uppermost Ordovician), Gunneklev in the Skien-Langesund District, Oslo Region, Norway.
A-B. Two views of the holotype (PMO 72154) C. A small specimen. All figures ab. 10 x.

It is found on the flattened side of the coralum. Except in very small specimens (fig. 2c) all the other septae are of equal strength.

The columellar structure is rather variable, consisting partly of a blade-like expansion from the central part of the cardinal septum, and partly of an irregular vesicular mass formed by coalecesence of a number of septa. In young specimens the former structure is dominating, in older ones more or less of the vesicular material is developed. In adult specimens the blade-like structure seems to be connected with the counter septum, but in the younger ones it formes an expansion of the central part of the cardinal septum.

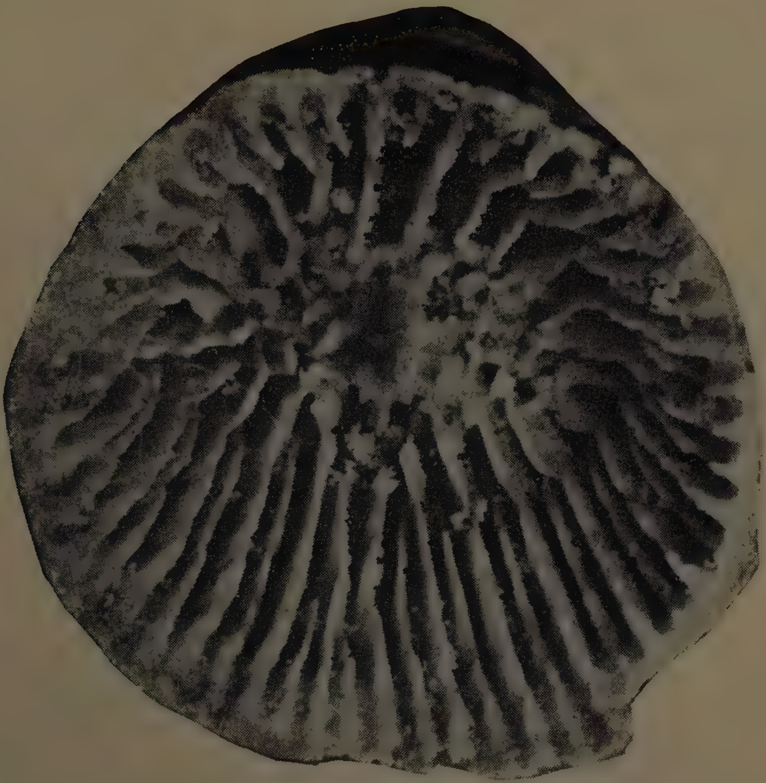


Fig. 3. *Holophragma duncanae* sp. n. from zone 5b (Uppermost Ordovician) at Gunneklev, in the Skien-Langesund District, Oslo Region, Norway.
A large specimen (PMO 72156) showing almost circular outline, columella consisting of the swollen cardinal (?) septum and vesicular tissue. Magnification ab. 10 x.

No diaphragms are observed.

Remarks: This species differ form the type species in the thinner septa, number of septa, and the structure of the columella. The development of the septa, and the shape is remarkably similar in the two. *H. calceoloides* is larger than *H. duncanae.*

The specimen recorded as *H. calceoloides* by SOSHKINA (1955, pl. VIII, figs. 6—7) from the Silurian of Podskamenskaja Tunguska seems to lack columella, and has a large, open cardinal fossula, and twisted septa. It appears, therefore to differ both from the species described here, and *H. calceoloides.*

According to DUNCAN (1957) most of the species referred to *Holophragma* from the Upper Ordovician of North America belong to the genus *Bighornia* DUNCAN, which differ from the present species and other *Holophragma's* in the absence of the cardinal septum, and in being flattened on the counter side instead of the cardinal one. This species is the first real *Holophragma* recorded from the European Ordovician. It is remarkable that both the Upper Ordovician and Lower Silurian has a number of different polygonal or other nonconical rugose corals (*Bighornia*, *Holophragma* and *Goniophyllum*). This might have some importance in the study of the phylogeny of the rugose corals, and because these forms generally are shortlived, for the stratigraphy.

The present find was immediately associated with the externally similar genus *Bighornia* because several members of the Red River fauna invaded NW-ern Europe in 5b-time, and the presence of that genus was expected (cf. DUNCAN 1957, p. 611). At present the coral fauna of the Upper Ordovician and Lower Silurian of the Oslo Region is not well enough known to allow any definite conclusions about the distribution of the calceoloid rugose corals. *Bighornia* and other genera might well be present beside *Holophragma*.

REFERENCES.

DUNCAN, H. 1956: Ordovician and Silurian Coral Faunas of Western United States. Bull. U. S. Geol. Survey *1021-F* pp. 209—235, pls. 21—27. Washington D. C.

— 1957: *Bighornia*, a new Ordovician Coral Genus. Journ. Paleont. *31*, pp. 607—615, pl. 70. Menasha. Wis.

LINDSTRÖM, G. 1896: Beschreibung einiger Obersilurischer Korallen aus der Insel Gotland. Bihang K. Svenska Vet.-Akad. Handl. *21*, afd. 4, No. 7. 50 pp., 8 pls. Stockholm.

SOSHKINA, E. D. 1955: Korallii *in* Materiali po Faune i Flore Paleozoii Sibiri. Trudi Paleont. Inst. Akad. Nauk SSSR *56*, 196 pp. 23 pls. Moskva.

OLIVER, W. A. JR. 1958: Signifiance of external form in some Onondagan rugose Corals. Journ. Paleont. *32*, pp. 815—837, pls. 104—106. Menasha, Wis.

Manuscript received July 7, 1960.
Printed March 1961.

NORSK GEOLOGISK FORENING

Møte nr. 389, torsdag 6. oktober 1960 kl. 19.

Drammensveien 78, Oslo.

Tilstede 38 medlemmer og 10 gjester.

Formannen ledet møtet.

Referat fra møte nr. 388 ble lest opp og godkjent.

Konservator Dons gav en kort orientering om den internasjonale geologiske union.

Følgende nye medlemmer ble innvalgt:

459. Statsgeolog dr. philos. Sigurd Hansen, Danmarks Geologiske Undersøkelse, Charlottenlund, Danmark. Foreslått av Kari Egede Larssen og H. Rosendahl.
460. Prof. dr. William B. N. Berry, Department of Paleontology, University of California. Berkley 4, California, U.S.A. Foreslått av G. Henningsmoen og N. Spjeldnes.
461. Dosent dr. Krztytzof Birkenmajor, Laboratory of Geology, Polish Academy of Science, Cracow, Polen.
462. Dr. Lincoln R. Page, Chief of New England geology branch of U.S. Geological Survey. Begge er foreslått av Natascha Heintz og T. Siggerud.
463. Dr. Anthony D. Wright, Department of Geology, The Queen's University Belfast, North-Ireland. Foreslått av G. Henningsmoen og N. Spjeldnæs.

Statsgeolog *Thor L. Sverdrup* holdt foredrag om «Granittpegmatitter i Tysfjord».

Etter foredraget deltok Gjeldsvik, Bjørlykke og foredragsholderen i ordskiftet.

Statsgeolog *Rolf Feyling-Hanssen* holdt foredrag om «Den senkvartære landhevning på Svalbard».

I ordskiftet etter foredraget deltok Strand, Øyen, Rosendahl, Bjørlykke og foredragsholderen.

Møte nr. 390, torsdag 3. november 1960 kl. 19.

Drammensveien 78, Oslo.

Tilstede 32 medlemmer og 23 gjester.
Viseformannen ledet møtet.
Referat fra møte nr. 389 ble lest opp og godkjent.

Følgende nye medlemmer ble innvalgt:

464. Dr. Leslie J. D. Fernando, Geological Survey and Mine Department, Colombo 2, Ceylon. Foreslått av Barth og Dons.
465. Donald M. Ramsay, Ph. D., Department of Geology, Queen's College, Dundee, Skotland.
466. Brian A. Sturt, Ph. D., Geology Department, Bedford College, Regent's Park, London N. W. 1. Foreslått av Barth og Heier.
467. Professor Charles F. Davidson, Department of Geology, The University, St. Andrews, Skotland. Foreslått av Dons og Neumann.
468. Professor W. Uytenbogaardt, Geologisch Instituut, Vrije Universitet, De Lairessestraat 142, Amsterdam, Nederland. Foreslått av Heier og Siggerud.
469. Dr. R. H. Filby, Geologisk Museum, Sarsgt. 1, Oslo.
470. T. G. Elder, Geologisk Museum, Sarsgt. 1, Oslo.
471. Miss M. L. Busé, 1a Av. 6—18 Zona 14, Guatemala City, Guatemala. Foreslått av Dons og Neumann.

Det referertes en skrivelse til foreningen fra Dons i egenskap av sekretær for Den norske komité for XXI Internasjonale Geologkongress. Komitéen ble oppnevnt i statsråd våren 1957 efter at foreningen hadde gitt forslag til dens sammensetning. Efter at kongressen nå er avviklet, gjenstår det å avslutte regnskapene. Regnskapene for ekskursjonene, som ble finansiert ved deltakernes avgifter, er for de aller flestes vedkommende blitt innsendt fra ekskursjonslederne, og foreningen ble anmodet om å oppnevne en revisor til å gjennomgå disse regnskaper. — Til å utføre dette revisjonsarbeide ble Hagemann valgt.

Statsgeolog *Fredrik Chr. Wolff* holdt et foredrag: «Verdalens geologi».

I ordskiftet efter foredraget deltok Dietrichson, Holtedahl, Spjeldnæs, Strand, Vokes og foredragsholderen.

Konservator dr. *Nils Spjeldnæs* holdt et foredrag: «Polflytning og kontinentalforskyvning».

I ordskiftet efter foredraget deltok Gjessing, Holtedahl, Høeg, Marthinussen, fru Ormestad, Rosendahl, Strand, Størmer og foredragsholderen.

Møte nr. 391, julemøte, torsdag 15. desember 1960 kl. 19.

Drammensveien 78, Oslo.

Tilstede 38 medlemmer og 22 gjester.
Referat fra møte 390 ble lest opp og godkjent.

Følgende nye medlemmer ble innvalgt:

472. Laboratorieingeniør Roar Solli, Norges geologiske undersøkelse, Geologisk museum, Tøyen. Foreslått av Thor L. Sverdrup og Harald Skålvoll.
473. Bergingeniør Magne Ødegård, Militærkontoret, Veidirektoratet, Schwensensgt. 3—5, Oslo. Foreslått av Neumann og Jøsang.

Deretter holdt førstekonservator *J. A. Dons* et kåseri: «Omkring geologkongressen og ekskursjonene», med lysbilder fra noen av ekskursjonene og kommentarer fra ekskursjonslederne.

Professor *Ivar Hessland* viste film fra Svalbard-ekskursjonen.

Det ble overrakt en gave fra N.G.F. til førstekonservator Dons som en takk for hans innsats for geologkongressen 1960, og en blomsterbukett til fru Dons i samme anledning.

Generalforsamling, torsdag 9. februar 1961 kl 19.

i Drammensveien 78, Oslo.

Tilstede var 33 medlemmer og 4 gjester.

Formannen ledet møtet. I Siggeruds fravær fungerte Kari Egede Larssen som sekretær.

Årsmelding for 1960 ble lest opp og godkjent. Årsmeldinger fra Bergens og Trondheims geologiske klubber ble referert.

Foreningens regnskap for 1960 ble godkjent. Revisorenes og generalforsamlingens henstilling fra forrige år om å anbringe foreningens midler på en mer hensiktsmessig måte enn som bankinnskudd er tatt til følge, idet de disponible midler, kr. 10 000,— er plasert i statsobligasjoner. Dette vil først komme til syne i regnskapet for 1961.

Gjelsvik ga rapport fra Komiteen for Berg og steinindustriens fond, de disponible midler, kr. 20 000,— er plasert i en fordelaktig pantobligasjon.

Valg:

Formann:	*J. A. Dons*
Viseformann:	*T. Strand.*
Sekretær:	*T. Siggerud.*
Redaktør:	*N.-H. Kolderup.*

Redaksjonskomite: *T. Barth, J. Bugge, A. Heintz, O. Holtedahl, T. Strand.*
Styremedlemmer: *S. Skjeseth, A. Eriksen* og formennene i Bergens og Trondheims geologiske klubber, *H. Holtedahl* og *Chr. Oftedahl.*
Revisorer: *G. Henningsmoen* og *K. Ø. Bryn* med *F. Hagemann* og *A. Hjelle* som varamenn.
Reuschmedaljekomite: *J. A. Dons, H. Neumann* og *A. Heintz.*

Som foreningens representanter til Sulitjelmafondets råd ble valgt *J. Bugge* og *S. Skjeseth* med *G. Henningsmoen* som varamann.

A. Granli ble anmodet om å fortsette som foreningens forretningsfører.

Reuschmedaljen for 1960 ble utdelt til dr.philos. *Knut S. Heier* for hans avhandling: The geology of the Ørsdalen district, Rogaland, S. Norway, Norsk Geol. Tidsskr. Bd. 36, s. 167—211, 1956.

Årsmelding for 1960.

Siden forrige generalforsamling er ett medlem avgått ved døden, lensmann P. O. Ottesen.

4 medlemmer er utmeldt av foreningen.
I samme tidsrom er det innvalgt 22 nye medlemmer.
Foreningen har nå 283 medlemmer, derav:
2 korresponderende,
138 livsvarige,
143 årsbetalende medlemmer.

Det er i siste år avholdt 6 ordinære møter, med et samlet fremmøte av 233 medlemmer og 94 gjester.

Av Norsk Geologisk Tidsskrift er utkommet Bd. 40, hefte 1—4.

Styret har i 1960 bestått av:
H. Bjørlykke, formann.
T. Strand, viseformann.
T. Siggerud, sekretær.
K. Egede Larssen, styremedlem.
G. Aasgaard, styremedlem.
Formennene i Bergens og Trondheims geologiske klubber,
H. Holtedahl,
og
Chr. Oftedahl, styremedlemmer.
A. Granli har virket som foreningens forretningsfører.

Utdrag av regnskap for det ordinære budsjett for 1960.

Inntekt

Beholdning, overført fra 1959	kr.	2 116,71
Innbetalt kontingent i 1960	-	1 020,00
Tilskudd fra staten (1/7—59 — 31/12—60)	-	1 800,00
— fra N.A.V.F. (1/7—31/12—60)	-	7 500,00
Abonn. og salg av tidsskriftet	-	8 917,33
Renter av bankinnskudd	-	74,23
— livsvarige medlemmers fond	-	950,87
— berg- og steinindustriens gavefond	-	346,87
	kr.	22 726,01

Utgift

Tidsskriftet:				
Trykking, Bind 39, hefte 4; Bind 40, hefte 2			kr.	13 273,00
Klisjeer, Bind 39, hefte 4; Bind 40, hefte 2			-	2 502,00
Korrektur og tegnearbeide			-	1 092,20
Ekspedisjon av tidsskriftet			-	983,70
Arbeidshjelp, porto og skrivesaker			-	565,00
Møter			-	503,03
Diverse			-	367,82
Beholdning, overført til 1961: I kasse.	kr.	551,05		
I bank .	-	2 888,03	-	3 439,08
			kr.	22 726,01

Livsvarige medlemmers fond 1960

				Inntekt		Utgift
1/1—1960. Fondets kapital	kr.	22 960,00				
Innbet. kont. i 1960	-	1 200,00				
	kr.	24 160,00				
Herav i statsobl. og aksjebrev	kr.	14 100,00				
I bank	-	10 060,00				
31/12—1960 Urørlig kapital	-	24 160,00				
Renter av verdipapirer			kr.	720,50		
— bankinnskudd			-	251,37		
Forvaltning av verdipapirer					kr.	21,00
Overført til det ordinære budsjett					-	950,87
			kr.	971,87	kr.	971,87

Berg- og steinindustriens gavefond.

Fondets kapital	kr.	20 000,00				
Renter av bankinnskudd			kr.	346,87		
Overført til ord.budsjett					kr.	346,87
			kr.	346,87	kr.	346,87

Regnskapet revidert av J. D. Dons og G. Henningsmoen, 12/1—1961.

Møte nr. 392, torsdag 9. februar 1961

i Drammensveien 78, Oslo

Til stede 33 medlemmer og 6 gjester.
Referat av møte nr. 391 ble lest opp og godkjent.

Følgende nye medlemmer ble innvalgt:

474. Verksmester John Martin Karlsen, A/S Grønseth & Co., postboks 185, Oslo. Foreslått av T. L. Sverdrup og O. M. Halfdansen.
475. Cand.mag. Ellen Sigmond Kildal, Geologisk institutt, Universitetet i Bergen. Foreslått av I. Hernes og I. Bryhni.
476. Stud.real. Frank Nikolaisen, Geologisk museum, Oslo NØ. Foreslått av N. Spjeldnæs og G. Henningsmoen.
477. Stud.real. Torgeir Falkum, Nyquistveien 21, Nordstrandshøgda. Foreslått av H. Neumann og J. A. Dons.
478. Mr. John H. MacDonald Whitaker, Department of Geology, Leicester University, England. Foreslått av N. Spjeldnes og G. Henningsmoen.

Dons ga en redegjørelse om Geologisk Union og planene for Norges opptagelse i denne.

Statsgeolog *Steinar Skjeseth* holdt deretter aftenens foredrag: «Resultater av diamantboringer i Femundtrakten».

I diskusjonen etter foredraget deltok Holtedahl, Strand, Vokes, Gjelsvik og foredragsholderen.

BERGENS GEOLOGISKE KLUBB

Årsmelding for 1960

Møte nr. 79, 15. februar.
Professor *Ivan Th. Rosenqvist:* Inntrykk fra en reise i Syd-Kaukasus.
14 tilhørere.

Møte nr. 80, 21. februar.
Vit. ass. *Inge Bryhni:* Trekk fra geologien i Eikefjord, Sunnfjord.
11 tilhørere.

Møte nr. 81, 11. mai.
Lektor *Isak Undås:* Den gamle jernvinna i Ustedalen.
12 tilhørere.

Møte nr. 82, 20. oktober.
Professor *Anders Kvale:* Litt nytt fra seksjonen om tektonikk ved kongressen i København.

Vit. ass. *Inge Bryhni:* Trekk fra diskusjonen om feltspater ved kongressen i København.
13 tilhørere.

Møte nr. 83, 26. oktober.
Utflukt til *A/S Norwegian Talc's* mølle på Knarrevik og laboratorium på Paradis.
26 deltakere.

Møte nr. 84, 23. november.
Cand. mag. *Kjell Tjølsen:* Trekk av geologien på Ombo, Rogaland.
Cand. mag. *Brit Løberg:* Geologien på kartblad Synnfjell.
16 tilhørere.

Møte nr. 85, 15. desember. Julemøte.
Vit. ass. *Inge Bryhni:* Gjennom ørkener og nasjonalparker i U.S.A. Med fargelysbilder.
Etter kåseriet festlig samvær i «Klubben», Studenthuset.
57 deltakere.

Styret var i 1960:

Formann:	Dosent Hans Holtedahl.
Viseformann:	Lektor Isak Undås.
Sekretær:	Vit. ass. Inge Brynhi.

Styret ble gjenvalgt for 1961.

TRONDHEIMS GEOLOGISK KLUBB

Årsmelding for 1960.

Møte nr. 21, 3. mars.
Professor dr. *Ivan Th. Rosenqvist:* Inntrykk fra en reise i Sovjet.
26 tilhørere.

Møte nr. 22, 30. mars.
Geolog *Tan:* Litt om Kaledonidene i Reisa-dalen.
17 tilhørere.
I diskusjonen etter fordraget deltok foruten foredragsholderen også Geis, Skjerlie, Wennervirta og Oftedahl.

Møte nr. 23, 20. oktober.
Bergingeniør *Brynjulf Dietrichson:* Om Jotuneruptivdekkene i det østlige Jotunheimens randområde og den generelle betydning av deres interpretasjon.
15 tilhørere.
I diskusjonen etter foredraget deltok foruten foredragsholderen også Bugge og Oftedahl.

Foredragsholderen refererte hovedtrekkene i sin avhandling i «Report of the International Geological Congress XXI Session Norden, *1960,* Part XIX., Caledonian Orogeny» pp. 64—88, presentert i Kb.havn 22/8 (også kort referert på Det nordiske geol. vintermøte i Uppsala 7/1 s. å.). Den bygger på der refererte arbeider, særlig av V. M. Goldschmidt, J. H. L. Vogt, Th. Vogt, forf. egne (1954—58) og hans før upubliserte resultater.

Møte nr. 24, 19. desember.
Julemøte med festlig samvær og lysbilder ved vit. ass. *O. Jøsang.* Tilstede 22 personer.

Styret i 1960 var:

Formann:	Chr. Oftedahl.
Sekretær:	O. Jøsang.
Styremedlem:	H. Carstens.

Styret ble gjenvalgt for 1961.

Vulkanologisk symposium i Italia

september 1961.

På kongressen til den Internasjonale Union for Geodesi og Geofysikk (IUGG) i Helsingfors august 1960 ble det i den vulkanologiske assosiasjon foreslått at det skulle arrangeres et internasjonalt symposium om ignimbritter og hyaloklastitter i Catania i annen halvdel av september 1961. Dette forslag ble godtatt av assosiasjonen og senere av sekretariatet i IUGG. Invitasjon og foreløpig program er nu kommet og programmet er som følger.

15. sept.	Ankomst Pisa.
15. »	Buss fra Pisa til Arcidosso. Ignimbritter og rheoignimbritter.
17. »	Monte Amiata med ignimbritter, lavastrømmer etc.
18. »	Besøk på dampkraftverk ved Monte Amiata og avreise til Rom med besøk i vulkanstrøket ved Latium.
19. »	Ekskursjon til Latiumvulkanen. Avreise til Catania.
20. »	Symposium over ignimbritter i det vulkanologiske institutt i Catania universitetet.
21. »	Fortsettelse av symposiet.
22. »	Ekskursjon til hyaloklastitter ved Monte Iblei.
23. »	Symposium om hyaloklastitter i Catania.
24. »	Ekskursjon til Etna og Acicastello.

Ekskursjonsgebyr er 30 dollar. Anmeldelsesfrist er 15. juni d. å. I tilslutning til selve symposiet vil det før dette bli arrangert en 3 dagers ekskursjon til Auvergne i Frankrike og en seks dagers ekskursjon etter symposiet til de Lipariske øer og Vesuv med Campi Flegrei. Det er adgang for alle profesjonelle geologer til å delta. Nærmere opplysninger og innmeldelsesskjema hos undertegnede.

Christoffer Oftedahl.

Den 12. kongress i den Internasjonale Union for Geodesi og Geofysikk (IUGG).

Kongressen ble avholdt i Helsingfors i dagene 25. juli til 5. august 1960. Det ble avholdt samtidige møter i alle de syv internasjonale assosiasjoner som unionen består av, nemlig assosiasjonene for geodesi, for seismologi og den indre jords fysikk, for meteorologi og atmosfærisk fysikk, for geomagnetisme og aeronomi, for fysisk oseanografi, for videnskapelig hydrologi og for vulkanologi. For geologer var det helt vesentlig vulkanologiassosiasjonens møter som hadde interesse, dernest var det mange geologer som gikk på endel foredrag i seismologi hvor det var med tydninger av jordskorpens og dens øvre mantels struktur, og oseanografi.

Møtene i vulkanologiassosiasjonen var gruppert i fire seksjoner etter hverandre, og disse seksjoner vil for fremtiden bli faste grupper innen assosiasjonen.

De er:

1. Aktive vulkaner.
2. Vulkaners fysikk.

3. Magmaenes fysikalske kjemi.
4. Paleovulkanologi og plutonisme.

Vulkanologiassosiasjonen omfatter således nesten hele fagfeltet til eruptivpetrologien. Nesten alle de fremførte foredrag var da også av den største interesse for geologer som arbeider i eruptivpetrologi, og relativt lite ble fremført av vulkanologi i egentligste og snevreste betydning. Både foredrag og etterfølgende diskusjoner i et auditorium på mellom 25 og 50 geologer fra hele verden var meget interessante og lærerike. Det kan bemerkes at assosiasjonens møter faktisk gav den besøkende mer enn den tilsvarende seksjon gjorde under Kjøbenhavner-kongressen. Det er således god grunn til å hevde at den vulkanologiske assosiasjon nærmest hører hjemme i en evt. geologisk union og ikke i den geofysiske.

Professor Alfred Rittmann som nu virker ved universitetet i Catania, ble gjenvalgt som president i den vulkanologiske assosiasjon for en ny 3-års periode. Professor Francesco Penta i Rom ble valgt til generalsekretær for assosiasjonen. Til slutt kan nevnes at professor Rittmann meddelte at det ved hans vulkanologiske institutt ved universitetet i Catania vil bli opprettet et vulkanologisk forskningssenter hvor man vil ta imot geologer fra alle land som ønsker å studere vulkanologi. Slikt studium vil da bli knyttet til Rittmanns studium av den aktive vulkan Etna og av de submarine vulkanske dannelser syd for Etna.

Christoffer Oftedahl.

CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 12. Hisingerite in "Dark" Larvikite.

By

R. V. DIETRICH

(Virginia Polytechnic Institute, Blacksburg, Virginia, U.S.A.)

Abstract. Magnesium- and calcium-bearing hisingerite, typically with extremely fine magnetite inclusions, occurs spatially associated with pyroxene (and very possibly as an alteration of olivine) in "dark" larvikite. It has a specific gravity of ca. 1.80, a waxy to subvitreous luster, is black to dark brown in reflected light and golden- to wine-brown by transmitted light, and has a greenish brown to olive green streak. Optical studies indicate the presence of intimately intermixed amorphous and crystalline phases. The amorphous phase has an index of refraction that ranges from 1.480 to 1.486. The crystalline phase has a beta index that ranges from 1.470 to 1.478, is biaxial negative, has a 2V of 60° to 70°, and has oblique extinction. *X*-ray diffraction patterns consist of three broad reflections with *d* values of their central portions of ca. 4.44, 2.575, and 1.535. Differential thermal analyses show the following reactions: large endothermic with peak at ca. 125° C., medium-sized exothermic with peak at ca. 450° C., and small exothermic at ca. 970° C. Chemical analyses, all of which have necessarily been made on contaminated material, are consistent with previously published analyses.

This study has pointed up the fact that relationships among materials called hisingerite, chloropal, nontronite, natural iron oxide gel., *etc.* need much further investigation and clarification. It is suggested that for a start the name hisingerite could be restricted to either the amorphous or the crystalline phase with the listed characteristics.

Introduction

Hisingerite was named by Berzelius (HISINGER, 1828) after W. Hisinger, his coworker during the early electrolysis experiments which led to Berzelius' famous electro-chemical theory. The type material came from Riddarhyttan in Vestmanland, Sweden.

Although it has been noted, by way of reference, that hisingerite was reported to occur at Solberg, Norway (DANA, 1892, p. 702), none was truly recorded from any Norwegian locality until the recent listing of two localities by NEUMANN (1959, p. 233)[1].

The hisingerite reported upon herein was discovered during the separation of minerals from each of the chief intrusive phases of the Oslo Series rocks in conjunction with the writer's current investigation of the possible utilization of K-feldspar obliquities as geological condition(s) indicators.

With these rocks, typically the lightest fraction consists chiefly of K-feldspar. However, the "dark" larvikite used proved exceptional. After the K-feldspar fraction was removed, a lighter-weight, nearly black, apparently magnetic fraction remained.

After numerous investigations it has been decided that the obtained data best support the conclusion that most of this material is hisingerite. This is a report of the determined data plus a discussion of certain aspects of them and their bearing on the use of the designation hisingerite.

Localities and host rock

The investigated material was first found to occur in one of the "dark" larvikites from the Mineralogisk Museum collection. This larvikite was collected by S. Bergstøl from the Klåstad Steinbrudd, Tjølling, in 1960. Subsequently, the material has been found also to occur in the only other "dark" larvikite checked — one collected by the writer, in 1959, at a quarry north of Viksfjord and 2—4 km east of Larvik (Valby).

Larvikite has been described petrographically by many workers (*e.g.*, BARTH, 1944) so will not be described in detail in this paper. Suffice it to say, the type in which the hisingerite was found to occur is the typically beautiful "dark" type. In thin sections of the Tjølling material this type can be seen to consist chiefly of oligoclase and alkali feldspar plus small amounts of varietal biotite, barkevikitic hornblende, and titanaugite and lesser amounts of accessory apatite, magnetite

---

[1] Apparently "i Nerike" was interpreted as "in Norway". The original report (Cleve and Nordenskiöld, 1866, p. 170) reads: "från Solbergsgrufvan i Nerike". Nerike is an old provincial designation for the area around Örebro in southcentral Sweden.

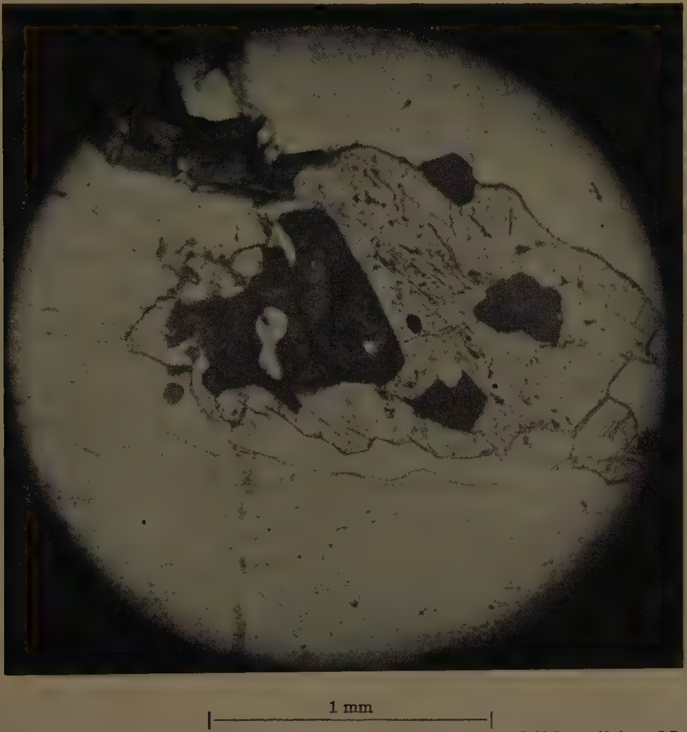


Figure 1. Hisingerite within titanaugite of Tjölling "dark" larvikite. Notice how shape is possibly suggestive of partially resorbed olivine.

and sphene plus traces (found to be about 0.01 per cent by weight) of the hisingerite.

The hisingerite has thus far been found to occur chiefly as subhedral to anhedral grains (probably pseudomorphous after olivine — see Figure 1) within pyroxene grains.

Previous work on hisingerite

As noted in the introduction, hisingerite was named by Berzelius. Subsequently, hisingerite has been described from numerous worldwidespread localities. Attention is called especially to the numerous references given by DANA (1892, p. 72), SUSTCHINSKY (1910), and HEWETT and SCHALLER (1925). Since the report of Hewett and Schaller, two papers concerning the mineral are particularly noteworthy — those of GORDON (1944) and SUDO and NAKAMURA (1952).

Actually, there is little agreement today as to either the definition or the place within the general mineral classification scheme for hisingerite. Commonly, identification has been based on chemical composition even though optical properties are inconsistent (see Table I).

Table 1.
Diverse optical data reported for hisingerite.

Locality:	Isotropy:	Index(es):	2V, sign, etc.	Reference:
Bellvue King, Idaho	iso-	1.44		Hewett & Schaller; Schaller
Wilmington, Delaware	iso-	1.45 —1.47		Hawkins & Shannon
Tjølling, Norway	aniso-	1.470—1.478	60°—70°,(–)	this paper
Tjølling, Norway	iso-	1.480—1.486		this paper
Riddarhyttan, Swed.	iso-	1.49 —1.53		Hewett & Schaller
not reported	iso-	1.49 —1.51, 1.59 & 1.66		Larsen
Parry Sound, Ontario	amorph. & crystalline	1.50 —1.56		Schwartz
Llallagua, Bolivia	iso-	1.52		Gordon
Various localities, e.g., Långban & Orijarvi, Sweden	aniso- (cryptocrystalline)	> 1.535	uniaxial (–) ??	Sustchinsky
Chognacota, Bolivia	iso-	1.57		Gordon
Wilmington, Delaware	aniso-	α—1.562 β—1.580 γ—1.582 β ranges — 1.552—1.595	small (–); parallel ext. (this is "canbyite")	Hawkins & Shannon
Minnie Moore, Idaho	aniso- & iso-	1.66	small (nearly uniaxial) (-) ?	Hewett & Schaller
Llallagua, Bolivia	aniso-	1.715—1.730	(–) elong.	Gordon
Edna May Deeps Mine, West Australia	aniso- & iso-	none given		Simpson
Kawayama, Japan	aniso- & iso-	none given		Sudô & Nakamura

Some workers have suggested tentatively that it is the iron-equivalent of allophane; others have suggested that it should be placed in the nontronite (and, thus, the montmorillonite) clan; *etc.; etc.* WINCHELL and WINCHELL (1951), without any stated basis, have classified it as a phyllosilicate, *i.e.*, made up of sheets of SiO_4 tetrahedra.

Origins suggested for the mineral are also diverse. Many occurrences appear to favor formation as the result of some weathering process(es). Relationships at other occurrences appear to preclude such an origin and to favor, instead, for example, formation as a hypogene mineral. The only consistency is that each suggested mode of formation makes the hisingerite secondary. On the basis of data thus far presented, all suggested origins appear to be quite tenable.

As mentioned, the only authentic report of hisingerite occurrences in Norway are those given by NEUMANN. He has recorded (1959, p. 233) the fact that there are specimens in the collection of the Mineralogisk Museum from "pegmatite dykes at Hyttåsen, Mjærum farm in the county of Östfold and at Drag, Tysfjord in the county of Nordland". Therefore, this paper presents the first descriptive material concerning hisingerite from Norway.

Physical properties

The chief physical properties of the mineral separated from the larvikite may be listed:

Color: dark brown to black in reflected light;
golden- to wine-brown by transmitted light.
Streak: olive green to greenish brown.
Luster: waxy to subvitreous.
Breakage: conchoidal fracture; one imperfect cleavage or parting (possibly inherited).
Diaphaneity: transparent in thin pieces.
Hardness: 3—3.5.
Specific Gravity: ca. 1.786.
Magnetism: *Non*-magnetic?

The low specific gravity is a noteworthy character of this hisingerite. As noted, it was because of this characteristic that the mineral was discovered. The given value was determined by Westphal-balance checking of the specific gravity of a fluid within which grains (-90 mesh) remained at any level at which they were placed. It is considered approximate because the fluid used contained much acetone which may have evaporated during the process of the measurements (thus making the liquid become heavier) and also because nearly all grains are contaminated by magnetite (S. G. — ca. 5.0). Nonetheless, this specific gravity is lower than the 2.5—3.0 range given for previously reported hisingerite. In fact, it is more nearly that generally reported for chloropal, *i.e.*, 1.727—2.105 (DANA, 1892, p. 701).

The "*Non*-magnetic?" needs explanation. It was discovered early in this investigation that a simple way to separate the mineral was by magnetic and heavyliquid methods. The mineral appeared to be magnetic. Actually, it now appears that it is not magnetic itself but reacts to a magnetic field because nearly all grains of it contain finely divided grains of magnetite. However, this is still subject to some question because a few grains with no microscopically (highest power with a petrographic microscope) visible inclusions of magnetite are slighhtly magnetic. It is thought that probably even these grains contain submicroscopic magnetite and that the apparent magnetic properties depend on this because a few similarly "pure-appearing" grains are *not* attracted by the same intensity magnetic field.

Optical properties

The following optical properties were determined on separated grains:

Crystalline and amorphous phases occur together in many individual grains.

Crystalline parts show: $\beta = 1.470 - 1.478$
Biaxial (—)
2V = 60° — 70° (variable)
Δ — low to moderate
Oblique extinction
Elongation (+)?
Color — golden- to wine-brown
Pleochroism — none

Amorphous phase shows: $n = 1.480 - 1.486$

Because the optical properties can be determined for the above designated crystalline part and because they are essentially the same for all of those parts checked, it would appear that this is a distinct crystalline phase and not merely a gel material with some parts made birefringent because of tension polarization as suggested by SUDÔ and NAKAMURA (*op. cit.*, p. 619). X-ray data are corroborative.

Thermal data

A tracing of the Differential Thermal Analysis curve for the hisingerite from the "dark" larvikite plus the curve for hisingerite contaminated by pyrite, as given by SUDO and NAKAMURA (*op. cit.*, p. 620), and a typical "diluted" pyrite curve comprise Figure 2.

The setup used in this investigation was an FH305 furnace in vertical position heated at the rate of 12.5° C/min.; Leeds and Northrup x—y recorder; Pt + Pt+ 10Rh thermocouple; three point system; scale multiplication of 2. The hisingerite was comprised by both the crystalline- and amorphous-appearing phases and was diluted approximately 1:1 with Al_2O_3 (curve A). Sudô and Nakamura also used a mixture — 1:3 = hisingerite:alumina and used a heating rate of 10° C/min. (curve B). The pyrite was also diluted (<20 per cent); its curve is taken from McLAUGHLIN (1957, p. 366) (curve C).

The exothermic reaction at ca. 465° C on curve B was attributed by Sudô and Nakamura to pyrite contamination. Curve A for the hisingerite from the "dark" larvikite is similar to their curve even to the point of also having an exothermic reaction at ca. 450° C. This Norwegian hisingerite has *no* pyrite contamination! Further, its only possible contaminants — feldspar and magnetite have no exothermic reactions near 450° C. (see, for example, data given by KÖHLER and WIEDEN, 1954 and MACKENZIE, 1957, pp. 305—306). It is true that pyrite may have an exothermic reaction at ca. 450° C. (see curve C). Possibly the double-hump on the exothermic "peak" of curve B is real and one of the "humps" belongs to pyrite whereas the other belongs to hisingerite. However, it appears more likely that the reported "less than 1%" pyrite contamination might not have given any recordable reaction and that the double hump "peak" is completely attributable to a hisingerite reaction.

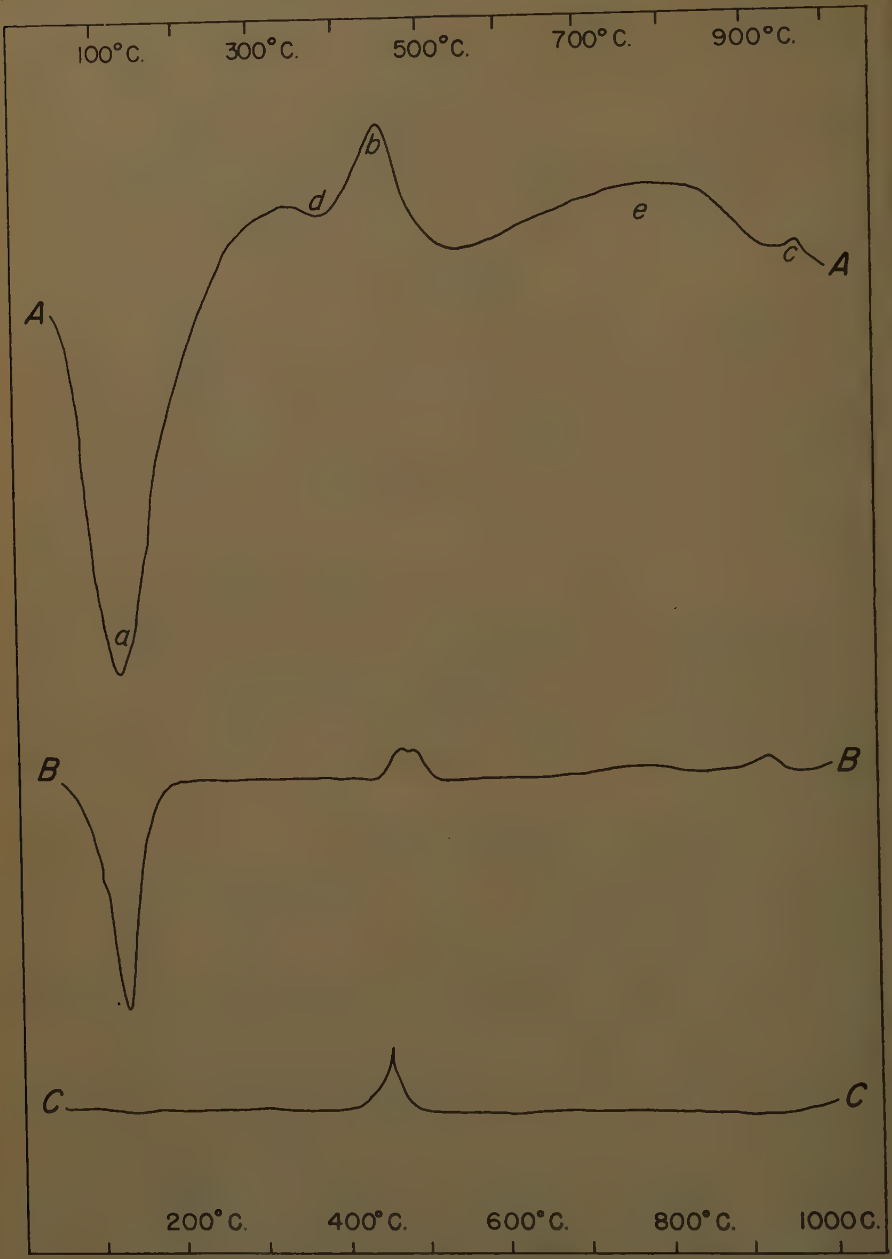


Figure 2. Differential Thermal Analyses Curves for A. Norwegian Hisingerite, B. Japanese Hisingerite, and C. Diluted pyrite. See text for full explanation.

If this is the case, the curve may be interpreted tentatively as follows: endothermic reaction (a) — loss of water; exothermic reaction at ca. 450° C. (b) — some structural adjustment, *e.g.*, (re)crystallization of all or part of dewatered (and amorphous ??) material; exothermic reaction at ca. 970° C. (c) — formation of something, *e.g.*, one of the high temperature modifications of SiO_2 or perhaps some spinel; the small endothermic reaction (d), if it actually is such, may, for example, represent loss of traces of water that were retained after the major endothermic reaction or perhaps some structural change such as breakdown of any retained crystallinity; the broad, more or less exothermic bulge (e) between the two better defined exothermic peaks may also be of little or no real significance — on the other hand, it may represent lag (re)crystallization of material not or not fully changed during the reaction manifest by the ca. 450° C. peak and/or some change in oxidation state or other reaction that occurs over a relatively wide temperature range.

The D. T. A. curve for hisingerite and the curve for a hydrated ferric oxide gel prepared by adding 9*N* alkali to 0.2*N* ferric chloride solution to pH 8.5 at 90° C. and washed free from salts after standing for two hours as given by MacKenzie (1957, p. 308) are also somewhat similar. However, this is the only ferric oxide gel, about which data are available, which has a D.T.A. curve with temperatures of reaction similar to those of hisingerite and, there is one marked difference — the lack of the exothermic peak at ca. 970° C. on the ferric oxide gel curve and its presence on the hisingerite curve.

X-Ray investigations

The hisingerite has been subjected to numerous *x*-ray diffraction setups. In each case, a powder was used. The following setups were employed: General Electric XRD-3 and XRD-5 diffraction units with spectrogoniometers and CoKα radiation with iron filter and CuKα radiation with nickel filter, respectively, were used at scanning rates of 1 degree 2θ/0.5 min. and 1 degree 2θ/5.0 min.; a Westinghouse unit with a 114.6 mm. camera and CuKα radiation with nickel filter was used with exposure times of 4—8 hours; and, a Norelco unit with a 6 mm camera and CrKα radiation with vanadium filter was used with exposure times of 6—24 hours on both untreated and glycolated hisingerite.

Only the CrK*a* radiation over 18 or more hours exposure times yielded patterns for anything other than the contaminants.

Three broad lines not attributable to any of the possible contaminants appear on all films exposed for 18 hours or more with the CrK*a* — vanadium filter setup. The *d* values for these broad reflections are given as Table II.

Also presented on Table II are the *d* values given by Sudô and Nakamura (*op. cit.*) and *d* values of highest intensity reflections for one of the nontronites investigated by Gruner (1935). The last-listed is presented to show that whereas all of the lines possibly "fit" nontronite, many of the strong nontronite lines are lacking in the pattern of the material from the larvikite.

Table 2.
X-ray data (numbers are *d* values in Å units).

Hisingerite in larvikite:			Hisingerite, Kawayama Mine, Japan (Sudô & Nakamura, op. dit.):	Nontronite, Farat-siho, Madagascar Gruner, 1935):*
edge (in):	edge (out):	ca. ave.:		
				13.9
				7.1
4.6668	4.2110	4.44	4.44	4.44
				3.54
2.6422	2.5109	2.575	2.61	2.588
				2.507
				1.72
				1.678
1.5430	1.5280	1.535	1.50	1.519
				1.481
				1.23

* Only relatively high-intensity reflections are given here.

Because there was such a partial "fit", however, the hisingerite was subjected to glycolation to see if it would exhibit any tendency to expand like nontronite (as a member of the montmorillonite clan) does. *No* change was evident in subsequently taken x-ray patterns.

Chemical analysis

The chemical analysis given as Table III represents the hisingerite from Valby *plus* contamination by magnetite. This magnetite contamination plus the fact that all iron is reported as Fe_2O_3 means that the reported Fe_2O_3 is too high. The silica value also must be at least slightly incorrect — it was determined optically that the "silica skeleton" which remained after solution by HCl (HEWETT and SCHALLER also refer to this — *op. cit.*, p. 33) was accompanied by about 20 per cent feldspar; this was taken into account, strictly on the basis of feldspar: silica ratios found to obtain as the result of grain counts made under the microscope, so far as establishing the value given in the analysis.

Table 3.

Approximate chemical analysis of hisingerite

SiO_2	36.67
Fe_2O_3	42.21
MgO	4.48
CaO	1.18
H_2O (> 110° C)	4.67
H_2O (< 110° C)	8.86
total	98.07

analysts: P.S. Dear & R.V. Dietrich

Spectrochemical analyses of the material show that it (again, *plus* magnetite contamination) contains, in addition to the reported major constituents, the following elements (with figures reported to the nearest numbers in the series .03, .015, .007, *etc.*): Be-.00015, Na-.03, Al-.3, Ti-.007, V-.0007, Cr-.0003, Mn-.7, Co-.007, Ni-.003, Cu-.007, Zn-.03, Sr-.0015, Ag-.000015, and Ba-.0007 (analyst: H.W. Worthing, U.S. Geol. Surv.).

General discussion and conclusions

It is quite evident that (at least on the basis of some previously published data) some of the reported characteristics of this hisingerite fit chloropal better than they do hisingerite (*e.g.*, the Specific Gravity). It is equally evident that published optical data for materials identified as hisingerite are inconsistent — especially the reported indices of refraction but even more particularly, so far as theoretical considerations are concerned, the admission of amorphous *and* crystalline materials as constituents of a single mineral species. Heretofore, only Sudô and Nakamura have published any *x*-ray or D.T.A. data for hisingerite. They noted that the D.T.A. data were identical for the two phases. Both their D.T.A. and *x*-ray data and those found to be characteristic of the material from Norway are in remarkable agreement.

Nonetheless, there is considerable confusion!

The present writer believes that much of this may be removed if hisingerite is defined strictly as either amorphous or crystalline. Although the former would appear to have priority, on the basis of the original description, the latter might be the better choice — optical, *x*-ray, and D.T.A. data which are coordinated and correlated to at least some of the material(s) called hisingerite are available for the crystalline phase. Thence, if this suggestion were followed, the so called amorphous phase, in particular, should receive further study (perhaps it is not "amorphous", for example, to electron diffraction) and perhaps be redefined as a separate species.

Until this is done, it seems best merely to call the material, the investigation of which is reported herein, simply hisingerite.

If this hisingerite is pseudomorphic after olivine, it is of replacement or alteration — *i.e.*, secondary — origin. If the Mg-content of the hisingerite reflects that of a preexisting olivine, that olivine would appear to have been a fayalite, probably near ferrohortonolite, in composition. Actually, however, BARTH (*op. cit.*, p. 78) has found the olivine in larvikite from "opposite Håö" to be a relatively Mg-rich olivine (Fo_{70}). This introduces the question — what was the composition of the olivine, if present, in the "dark" larvikite in which the hisingerite now occurs? Two alternatives and likely dependent resolutions of the problem of hisingerite genesis are: 1. If the olivine was of composition similar to that determined by Barth for the Håö larvikite, it would appear that ferric iron must have replaced most of the magnesium (which, of course, would have to have been taken away). As Professor Barth (personal communication, 1961) has noted with respect to this alternative "much more 'unlikely' replacement processes take place in pseudomorphism". Further, if the active solutions were high in ferric iron and low in magnesium, the hisingerite could have been precipitated (perhaps even with the accompaniment of magnetite formation) and magnesium could have been dissolved thus precluding serpentinization. *or* 2. If the olivine was an Fe-rich variety, unlike that found to occur in the Håö rock, the high Fe:Mg ratio might, in itself, account for formation of hisingerite in lieu of serpentinization plus formation of magnetite, (generally the process of alteration of the more common Mg-rich olivines). In any case, either of these alternatives appears more likely than any simple conditional control because of the presence of magnetite intimately intermixed with the hisingerite just as it typically is with serpentine after olivine. So far as these considerations are concerned, it will be most interesting to see if Fe-rich olivines from any localities have been altered to hisingerite.

Acknowledgements

Fru B. Nilssen aided with the separations made at the Mineralogisk Museum. Certain parts of the investigation were carried on in the V.P.I. laboratories of E. E. Barton, Jr., P. S. Dear, C. I. Rich, W. Richardson, and H. D. Ussery — each of these colleagues aided me in their respective laboratories. Miss J. J. Glass of the U.S. Geological Survey checked the optical determinations. I gratefully acknowledge these contributions.

REFERENCES CITED

Barth, T. F. W., 1944, Studies on the igneous rock complex of the Oslo region, II. Systematic petrography of the plutonic rocks: Skrf. Norsk Vid.-Akad., Oslo. I Mat.-Naturv., Kl. 9, pp. 76—83.

Cleve, P. T., and Nordenskiöld, A. E., 1866, Om sammensättningen af jern haltiga kolloïd-silikater: Öfversigt. Kungl. Vetenskaps-Akad. Förh., N:o 7, pp. 169—183.

Dana, E. S., 1892, The system of mineralogy of James Dwight Dana. . .etc. (6th edit.): Wiley, New York, N.Y., pp. 701—704.

Gordon, S. G., 1944, The mineralogy of the mines of the Cerro de Llallagua, Bolivia, Proc. Acad. Sci. Phila., Vol. XCVI, p. 355.

Gruner, J. W., 1935, The structural relationship of nontronites and montmorillonites: Am. Min., vol. 20, pp. 475—483.

Hawkins, A. C. and Shannon, E. N., 1924, Canbyite, a new mineral: Am. Min., vol. 9, pp. 1—5.

Hewett, D. F. and Schaller, W. T., 1925, Hisingerite from Blaine Co., Idaho: Am. Jour. Sci., vol. 10, 5th ser., pp. 29—38.

Hisinger, W., 1828, Analyse des mit den Namen Hisingerit Belegten Eisensilicats: Pogg. Ann. der Physik u. Chem., vol. 13, pp. 501—508.

Holdridge, D. A. and Vaughan, F., 1957, The kaolin minerals (kandites): in The Differential Thermal Investigation of Clays (Ed. by R. C. MacKenzie), Central Press, Aberdeen, Scotland, pp. 98—139.

Köhler, A. and Wieden, P., 1954, Vorläufige Versuche in der Feldspatgruppe mittels der D.T.A.: Neues Jb. Miner., Mh., H. 11, pp. 249—252.

Larsen, E. S., 1934, The microscopic determination of the nonopaque minerals (2nd edit.) USGS bull. 848, p. 266.

MacKenzie, R. C., 1957, The oxides of iron, aluminium, and manganese: in the Differential Thermal Investigation of Clays (Ed. by R. C. MacKenzie) Central Press, Aberdeen, Scotland, pp. 299—328.

McLaughlin, R. J. W., 1957, Other minerals: in the Differential Thermal Investigation of Clays (Ed. by R. C. Mackenzie), Central Press, Aberdeen, Scotland, pp. 364—388.

Neumann, H., 1959, Contributions to the mineralogy of Norway: No. 1 An Introduction: Norsk. Geol. Tids., Bd. 39, pp. 231—236.

Oftedahl, C., 1948, Studies on the igneous rock complex of the Oslo region, IX. The feldspars: Skrf. Norsk Vid.-Akad., Oslo. I Mat.-Naturv., Kl. 3, pp. 7, 18, 19, 48—49.

Robertson, R. H. S. and Twedily, A. E., 1953, Biogeochemistry of Skye diatomite: Clay Min. Bull., vol. 2, pp. 7—16.

Schaller, W. T., 1926, Index of refraction of hisingerite — a correction: Am. Jour. Sci., 5th Ser., vol. 11, p. 376.

Schwartz, G. M., 1924, On the nature and origin of hisingerite from Parry Sound, Ontario: Am. Min., vol. 9, pp. 141—144.

Simpson, E. S., 1920, Hisingerite: Jour. and Proc. Roy. Soc. West Australia, vol. 5, pp. 95—97.

Sudo, T. and Nakamura, T., 1952, Hisingerite from Japan: Am. Min., vol. 37, pp. 618—621.

Sustchinsky, P. P., 1910, Über den Hisingerite: Zeit. Krist. Min., vol. 47, pp. 231—237.

Winchell, A. N. and Winchell, H., 1951, Elements of optical mineralogy, Part II. Descriptions of minerals (4th edit.): Wiley, New York, N.Y., p. 400.

Manuscript received January 25, 1961.

Printed December 1961.

LAYERED GABBRO, HORNBLENDITE, CARBONATITE AND NEPHELINE SYENITE ON STJERNØY, NORTH NORWAY

By

Knut S. Heier*

Abstract. The geology of a small area on the island of Stjernøy, within the Seiland petrographic province in North Norway, is described. The rock types range from ultrabasic to nepheline syenite and pegmatite. Rocks of gabbroic affinities dominate.

An interesting relationship exists between layered gabbro, hornblendite, carbonatite, and nepheline syenite.

Introduction and acknowledgements

This report is concerned with the geology of a small part of Stjernøy, West Finnmark, North Norway.

The field work was done during 2 months in the summer of 1959 and the laboratory work was done in the Mineralogisk-Geologisk Museum, Oslo, in 1959 and 1960.

The work was started on the recommendation of Professor T. F. W. Barth, Mineralogisk Museum, Oslo, who has spent several summers in the general area of the Seiland province, and who has expert knowledge of the intricate genetical problems exhibited by the exposed rocks.

The work was planned as part of a large scale geological investigation of the Seiland province.

In the summer of 1959 Dr. E. Weedon, of Glasgow University, began a detailed study of the Bumansfjord layered gabbro complex on Seiland. Dr. Sturt, of London University, and Dr. Ramsay, of the University of St. Andrews, Dundee, at the same time began an investigation on Sørøy of the contact relations between gabbro and metamorphic supracrustal rocks. Results of these studies will be published later.

---

* Dept. of Geology, Rice University, Houston, Texas.

Mr. J. Hysingjord has made a general geological map of the western half of Stjernøy, including the area described in this report. The work of Hysingjord is available as a thesis for the degree of cand. real. at the University of Oslo (Hysingjord, 1960) and will be published later. I want to thank him for most generously allowing me to include some of his analytical data in this report. The existence of a regional geological map greatly accelerated the field work.

Nansenfondet is heartily thanked for a generous grant covering all expences connected with the field work. Christiania Spigerverk A/S, Oslo, gave some financial aid.

Location, topography and climate

Stjernøy is located about 50 km southwest of Hammerfest, West Finnmark, North Norway, at about 22—23°E longitude and 70—71°N latitude.

The topography is very rugged with narrow valleys and steep, sometimes inaccessible mountain summits attaining an elevation of about 1000 m.

Vegetation is restricted to the shore, with only moss and lichen found higher up. This, however, is enough to feed the reindeers which inhabit the island from May to October. The carbonatite on Stjernøy yields excellent soil on which grass is to be found several hundred metres above sea level.

The climate is cold and wet with frequent storms during autumn and winter. In the summer the sun stays above the horizon 24 hrs. resulting in somewhat higher temperatures, although summer snow is by no means rare. The summer of 1959 was especially cold and wet.

Habitation is restricted to a dozen families living close to the sea. Fishing is their major occupation. At Sørfjord there is a small shop visited by a local steamer three times a week.

Earlier geological investigations

Very little about the geology on Stjernøy has so far been published. Strand (1952) described the carbonatite and (1950) commented upon the weathering and glaciation phenomena. UNDÅS (1938) measured old shorelines and terraces.

The files of the Geological Survey of Norway contain reports of short visits to Stjernøy by KVALE and NEUMANN, 1938; POULSON, 1948; RAMBERG and SIGGERUD, 1953.

M. G. OOSTEROM mapped the eastern part of Stjernøy during the summer of 1954, and J. HYSINGJORD (1960) the western part.

General geology of the area, and scope of present investigation

Stjernøy belongs to a petrographic province of dominately mafic and ultramafic rocks on the northwest coast of Norway known as the Seiland province, Fig. 1. It is believed to be part of the Norwegian Caledonides. However, rocks of nonorogenic type; i.e., carbonatite and nepheline syenite are present and characteristic. U—Pb analyses of zircons from the nepheline pegmatites in the area are too unreliable (Table 1) to be of any value for age determinations. Even some of the youngest rocks in the area are affected by folding, and post Caledonian age appears unlikely.

Table 1. Radioactive age of minerals from the Seiland province (compiled from Neumann (1960).

Rock type	Mineral	Method	Age	Locality	Laboratory
Plumasite pegmatite . .	Zircon	Pb *-a*	$279^{+10}_{\div 30}$	Seiland	Oxford
Pegmatite ?	Zircon	Pb *-a*	$294^{+15}_{\div 40}$	Seiland	Oxford
Canadite pegmatite . .	Zircon	Pb *-a*	$366^{+20}_{\div 40}$	Seiland	Oxford
Pegmatite ?	Zircon	Pb *-a*	520	Stjernøy	U.S.G.S.
Canadite pegmatite . .	Zircon	Pb *-a*	910	Skarvvann Seiland	U.S.G.S.
Ringit pegmatite . .	Zircon	Pb *-a*	930	Skarvberg Seiland	U.S.G.S.

Parts of the Seiland province in areas adjacent to Stjernøy have been described by BARTH (1927, 1953), KRAUSKOPF (1954), OOSTEROM (1954, 1956), and PETTERSEN (1874 a, b, 1883). Most of these authors

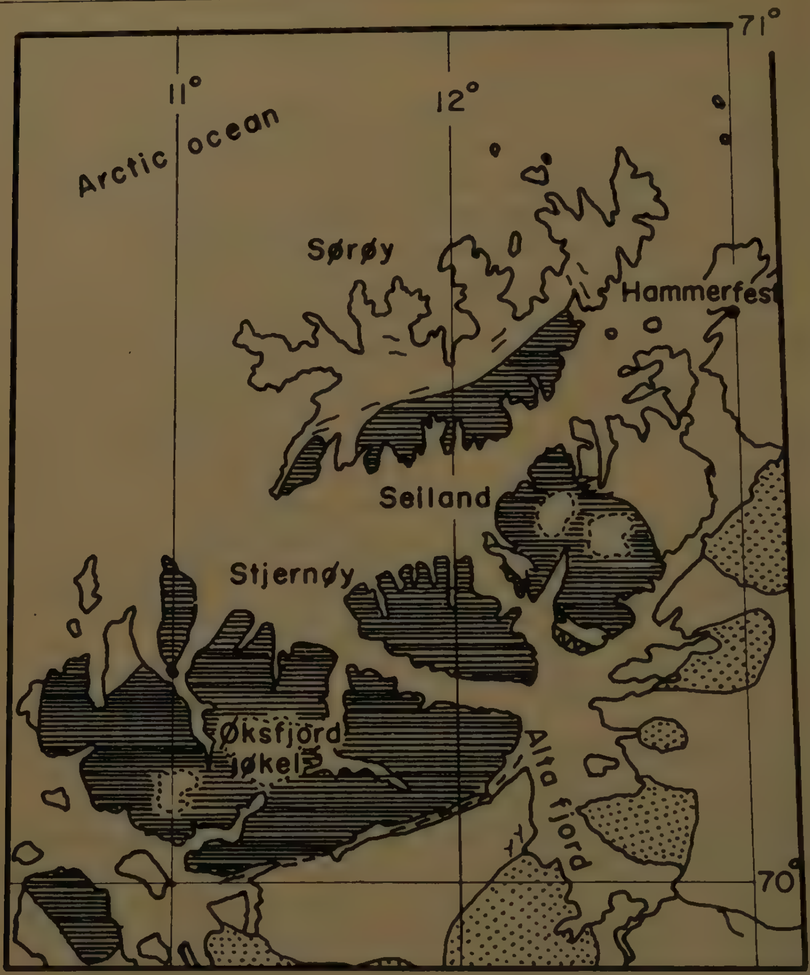


Fig. 1. The Seiland petrographic province. Horizontal lines = Gabbroic rocks, partly layered. White = Caledonian sediments. Dots = Raipas Formation. (Reproduced from Barth, 1953).

discuss very specific petrological problems but Krauskopf (1954) gives a regional description of rocks and structures of the Øksfjord area south and southwest of Stjernøy.

Gabbro is the characteristic and most abundant rock type within the Seiland province. Ultramafic rocks, anorthosite, amphibolite-hornblendite, garnet-biotite gneiss, syenite, nepheline syenite, ne-

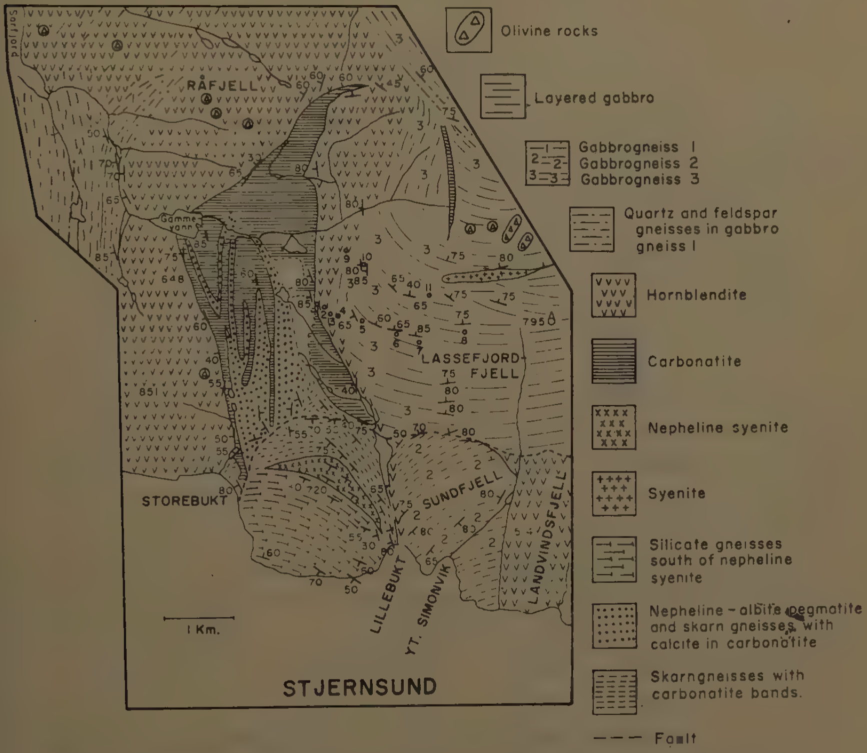


Fig. 2. Geological map of the part of Stjernøy discussed in this paper.

pheline albite (biotite) pegmatite, albite (biotite) pegmatite, carbonatite, and metalimestone occur in smaller amounts. All of these rock types (except metalimestone, and anorthosite which is found only as bands in banded gabbro) occur on Stjernøy and occur within the small area described in this report, Fig. 2.

The purpose of this study was to examine the carbonatite and nepheline syenite and their relation to the surrounding rocks.

In the Seiland province carbonatites are restricted to Stjernøy. Nepheline-bearing pegmatites occur on Stjernøy, Seiland, and in the Øksfjord area. So far massive nepheline syenite of intrusion like form is only known from Stjernøy. (Nepheline syenitic gneisses occur on Sørøy, Ramsay and Sturt, 1960, personal communication).

In the following descriptions and discussions reference is made to the map, Fig. 2.

Petrography

(a) The gabbros

Rocks which chemically, mineralogically, and texturally should be termed gabbros (or meta gabbros and gabbro gneisses) are subdivided into.

(a 1) Gabbro gneiss 1,
(a 2) Gabbro gneiss 2,
(a 3) Layered or banded gabbro,
(a 4) Gabbro gneiss 3.

(a 1) Gabbro gneiss 1, with associated syenitic and quartz rich gneisses. This rock type corresponds to the gabbro gneiss 1 of Hysingjord (1960), and probably also to the gabbro gneiss 1 of Krauskopf (1954), It occupies the far west area in Fig. 2 and continues beyond the map further west and north.

The gabbro gneiss sometimes shows a tendency towards layering of alternating light and dark bands. However, this tendency is not very pronounced. The light, plagioclase rich bands typically contain garnets.

Syenitic and quartz-rich gneisses occur as layers parallel to the foliation of the gabbro gneiss especially along the contact with the hornblendite to the east. Crosscutting contacts are never seen. Like the gabbro they are typically foliated. The quartz content is variable, and the rocks grade from near quartzites to syenites, monzonites and diorites. Garnets and hypersthene are typical constituents of these rocks. These gneisses are described as a part of the gabbro gneiss.

The following minerals are found in the gabbro: plagioclase, hypersthene, clinopyroxene, hornblende, (biotite), (green spinel), apatite, opaques.

The plagioclase composition is rather constant between An 50—55.

Orthopyroxene (hypersthene) may be the dominant femic mineral (16% of one rock counted from thin section). It typically shows varying stages of alteration into hornblende. North of Sørfjorddalen hornblende becomes gradually more important and the gabbro gneiss changes into hornblende gabbro and finally into hornblendite. In its last stage of alteration hypersthene is only found as small remnants in hornblende.

Table 2. Mode of gabbroic rocks within the gabbro gneiss 1 series (point counted thin sections).

	18/59 %	1/59 %	150/59 %
Plagioclase (An 50—55)	59	50	60
Hypersthene	16	7	tr.
Clinopyroxene	6	15	6
Hornblende	14	21	31
Biotite	tr.	tr.	—
Spinel	tr.	—	—
Apatite	—	tr.	—
Opaques	5	7	3

Clinopyroxene is a characteristic mineral present in varying amounts. As is the case with hypersthene, clinopyroxene is often observed altering into hornblende.

Hornblende is a major mineral. It occurs as independent grains as well as surrounding and replacing the pyroxenes. Whether or not it is always a secondary mineral is difficult to ascertain. Hornblendes of the two modes of occurrence are optically similar; both show pleochroism x,-light yellow; y, -light brown; z, -light brown.

Apatite and opaques are always present as accessory minerals, while biotite and green spinel may or may not be present.

The modes counted from thin sections of three of the rocks are given in Table 2.

The gabbro gneiss 1 with associated leucocratic layers is considered as a meta supracrustal (lavas, tuffs, sediments) series of rocks.

(a 2) *Gabbro gneiss 2.*

This rock type covers the area over Sundfjell. Hysingjord (1960) did not distinguish between gabbro gneiss 2 and 3. The two rocks may possibly be genetically related. However, there are certain petrographic differences between them, and their foliation directions are at an angle with each other.

The contact between the two gabbro series is shown in the topography as a narrow depression, marked with a heavy dashed line on Fig. 2. This depression also marks the border between the hornblendite north of Simonvika and the Lassefjordfjell gabbro (layered gabbro, a 3). It is explained as a fault line. The fault plane strikes east-west and dips steeply towards the south. The fault apparently continues across the carbonatite just north of the nepheline syenite (heavy dashed line on the map, Fig. 2). The presence of this fault explains the intensive mylonitization and the irregular strike and dip directions of the skarn gneisses north of the nepheline syenite. The fault also explains the angular disconformity between the banding of gabbro gneiss 2 and the layered gabbro (a 3).

Close to the fault the gabbro gneiss 2 exhibits fine-scaled banding, but in general the banding is not well developed.

The characteristic mineralogic feature of this gabbro gneiss type is orthopyroxene with beautifully developed green spinel symplectite. The spinel symplectite is restricted to the orthopyroxene and is not seen coexisting with, or even contacting, clinopyroxene. However, green spinel may occur as an independent (primary) mineral, often inside, and penetrating into clinopyroxene and hornblende. Whether the spinel symplectite in the orthopyroxene is the result of exsolution or is caused by simultaneous crystallization is difficult to decide on the basis of the textural evidence alone.

The plagioclase is An 65—70, and may be strongly secondarily altered. This alteration of the gabbro increases rapidly towards the west on approaching the carbonatite area.

In places the gabbro gneiss is mylonitized and contains zones of fine grained mineral aggregates. In these areas the larger plagioclase grains are bent and twisted and accompanied by irregular recrystallization of albite. Scapolitization and saussuritization of the plagioclase is also characteristic towards the carbonatite contact. Hypersthene and spinel disappear, while clinopyroxene, hornblende and biotite remain. At the same time the banding becomes less distinct and finally disappears completely.

In the southern part of this gabbro gneiss banding is nowhere developed but the rock is distinctly foliated. This foliation is parallel to the fine-scaled banding in the north. The typical minerals in the southern part of the gabbro gneiss are: plagioclase (An 60—70); scapolite (often as fairly large crystals inside plagioclase, this scapolite has a high birefringence which shows it to be rich in CO_2); calcite (as small grains often inside scapolite); clinopyroxene (partly altered into hornblende, the hornblende also occurs as independent grains);

biotite (colourless-redbrown); apatite (common, often as large crystals, sometimes it appears as if apatite veins penetrate the rocks); and opaques.

The texture of the gabbro gneiss 2 is typically mosaic, and its metamorphic nature cannot be doubted.

(a 3 and a 4) *Layered gabbro and gabbro gneiss 3.*

The layered gabbro is most characteristically developed on Lassefjordfjell where it forms the western continuation of a large area of similar rocks. On Lassefjordfjell the layered gabbro extends about 2 km from south to north (across the layering).

The layering is produced by a concentration of dark minerals in streaks and bands, often with indistinct and gradual contacts with the adjacent leucocratic bands. The thickness of the individual bands varies from very thin (approx. 1/2 cm) to several metres. The bands may be traced for hundreds of metres in the direction of the strike.

The dark bands only rarely attain the same thickness as the leucocratic bands. They are seldom more than 10—20 cm thick. When the dark bands occasionally become thicker they generally contain narrow plagioclase bands.

In spite of the great interest in layered gabbroic rocks, especially stimulated by the studies of Wager and coworkers on the Skaergaard layered complex on East Greenland, a systematic study of the layered gabbro on Stjernøy was not undertaken. This was not done because the layered gabbro complex on Seiland, which genetically probably is identical to the one on Stjernøy, is now being investigated in detail by Dr. Weedon of Glasgow University. For the purpose of the present study, however, the contact relations between the gabbro and the carbonatite, and the changes in the gabbro on approaching this contact, were investigated in some detail.

Concerning the layered gabbro on Seiland BARTH (1953, p. 195) mentions that in at least one respect it is different from the layered gabbro of the Skaergaard intrusion; e.g., the Seiland gabbro "has no definite side walls, but seems to extend without a clear break into the contiguous and analogously layered amphibolite — gneiss complex."

In a general way this also seems to be true of the layered gabbro on Stjernøy. (As mentioned before both the gabbro gneisses 1 and 2

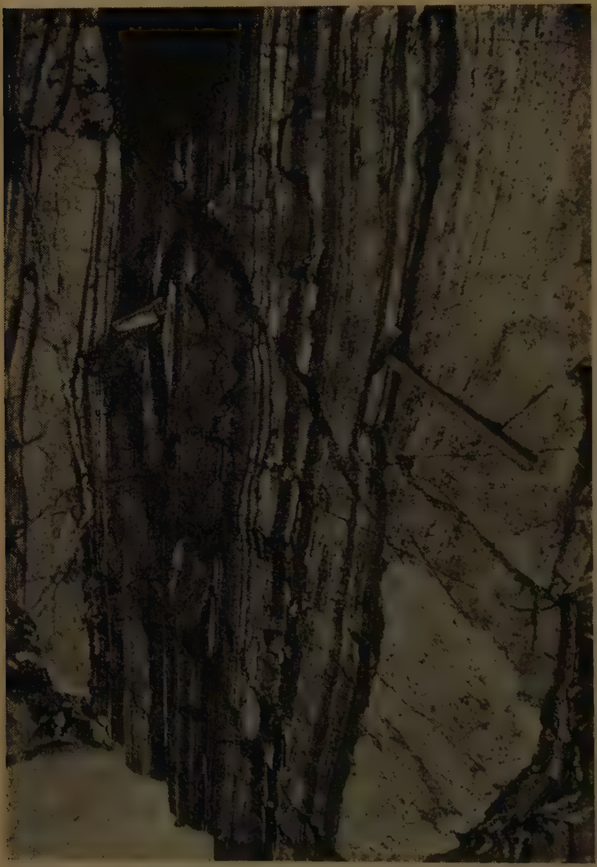


Fig. 3. Gradational change of the layered gabbro from East (Fig. 3) towards West (Fig. 5).

show indication of layering). However, a fault separates the layered gabbro from the gabbro gneiss 2 (and the hornblendite) to the south. The topography of the area also suggests a fault separating the layered gabbro from the gabbro gneiss to the north. Towards the carbonatite in the west, the layered gabbro grades into a gabbro gneiss (gabbro gneiss 3) in which the banding is not as well developed or is even non-existent. Hysingjord (1960) considered the layered gabbro and the gabbro gneiss 3 as unconnected, and stated that the layered gabbro fingers into the gabbro gneiss, and that the rocks of the border zone are tectonized.

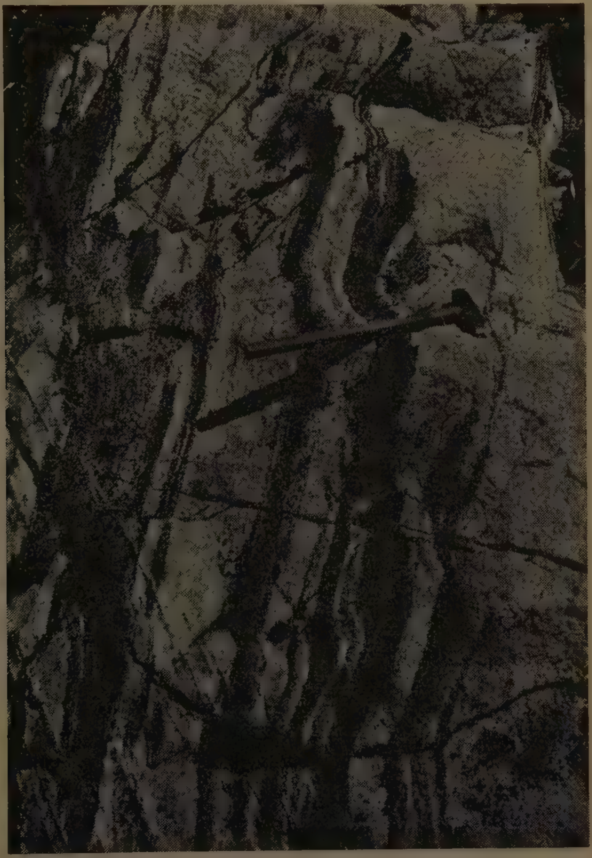


Fig. 4.
Gradational change of the layered gabbro from East (Fig. 3) towards West (Fig. 5).

The layered gabbro shows up well on aerial photographs, and the impresssion of "fingering" into the gabbro gneiss 3 is distinct. The change of strike (= direction of the bands) from east-west in the layered gabbro to north-south in the gabbro gneiss (Fig. 2) can also be seen to take place over a very short distance. However, a closer study of the photographs reveals a diffuse banding of the gabbro gneiss with a continuous transition into the layered gabbro. This is further clearly confirmed from field observations which also show that the sudden change in strike is continuous. Figs. 3, 4, and 5 show how the banding (layering) of the gabbro gradually becomes

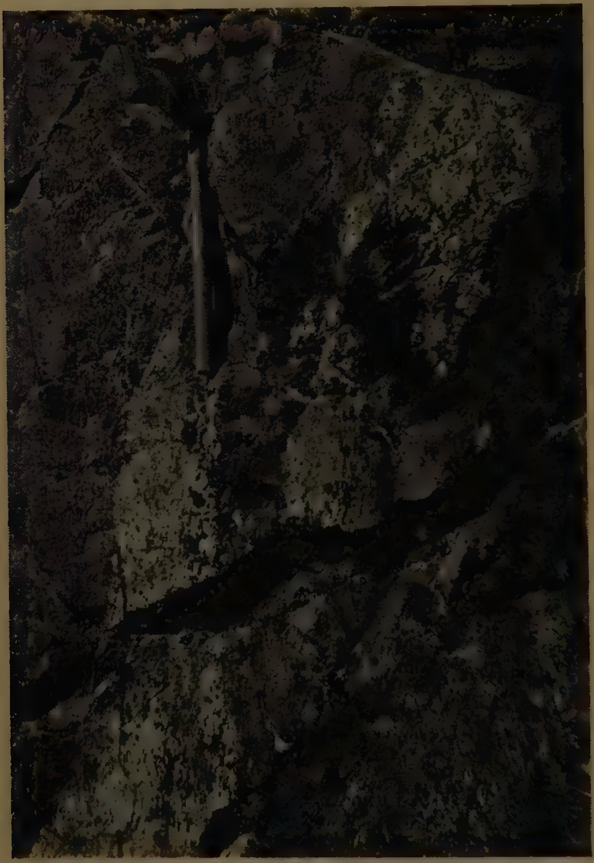


Fig. 5.
Gradational change of the layered gabbro from East (Fig. 3) towards West (Fig. 5).

more diffuse on approaching the carbonatite. The field impression is that of metasomatic growth of large hornblende crystals into the anorthositic bands; thus imparting to the rock a rather massive look.

On the south, at the carbonatite contact, the gabbro is a pyroxenite — hornblendite (1, 2, 3, Table 3) which towards the north grades into hornblendite with only little pyroxene.

In order to check the relations between the layered gabbro and the gabbro gneiss 3, two profiles were made from the carbonatite into the layered gabbro complex (Table 3, Fig. 2).

On Lassefjordfjell the banding is very well developed (Fig. 3).

Table 3. Modal composition of rocks along two prophiles from the carbonatite into the layered gabbro (marked with open circles and corresponding numbers on the map Fig. 2.)

Mineral	1	2	3	4 (dark)	4 (light)	4	5	6
Plagioclase	—	—	—	19.8	28.8	27.2	52.5	64.7
% An				35	35	50	59	65
Calcite	0.6	tr.	0.4	—	—	0.7	—	—
Scapolite	—	—	—	—	—	2.6	—	—
Orthopy.	—	—	—	—	—	—	—	—
Clinopy.	52.1	66.8	79.4	11.3	4.9	19.5	28.1	20.7
Hornbl.	18.6	20.9	8.7	54.7	55.8	42.0	10.9	13.7
Biotite	—	—	—	tr.	—	—	—	tr.
Green spinel	—	—	—	—	—	—	tr.	tr.
Opaques	19.7	8.7	7.8	8.3	5.2	5.4	8.5	0.9
Sphene	4.4	1.7	0.5	—	—	—	—	—
Apatite	4.6	1.9	3.2	5.9	5.3	2.6	tr.	—

Mineral	7 (light)	7 (dark)	8	9	10	11 (light)	11 (dark)
Plagioclase	67.1	41.5	37.3	—	13.5*	44.7	34.0
% An	60	65	70-75			55	70
Calcite	—	—	—	—	—	—	—
Scapolite	—	—	—	—	—	—	—
Orthopy.	—	tr.	—	—	—	—	—
Clinopy.	13.5	29.6	43.3	45.2	29.4	26.6	42.4
Hornbl.	15.2	16.3	14.6	42.6	53.8	18.9	11.8
Biotite	—	—	—	tr.	—	tr.	—
Green spinel	0.2	0.6	1.1	—	—	tr.	tr.
Opaques	4.0	12.0	3.1	6.4	2.2	9.8	11.8
Sphene	—	—	—	—	—	—	—
Apatite	tr.	—	—	5.8	1.1	—	—

* strongly saussuritized, new crystallization of albite without twin lamellae. 1—8, prophile 1; 1—3, pyroxenite-hornblendite; 4—8, gabbro gneiss — layered gabbro. 9—11, prophile 2.

The contacts between adjacent bands may or may not be sharp. The minerals in two adjacent dark and light bands are essentially identical but their relative proportions differ. The dark bands are essentially pyroxenitic and the light bands anorthositic. However, pure pyroxenites and anorthosites are rare. The bands are intermediate between the extremes.

The essential minerals in both types of bands are: plagioclase, clinopyroxene, hornblende, and iron ore (opaques). Some of the bands may consist essentially of iron oxides. Orthopyroxene is characteristically absent (only observed as traces in one thin section), and HYSINGJORD (1960) observed olivine in only one band. The plagioclase is a basic labradorite (An 60—70). HYSINGJORD (1960) found no systematic variation of the An content in a north-south profile across the layered gabbro. However, on approaching the carbonatite the plagioclase gradually becomes more albitic (Table 3) until it finally disappears.

The clinopyroxene is seen to be in various stages of alteration to hornblende. This is characteristic even in the central parts of the layered gabbro complex. The pyroxene/hornblende ratio decreases towards the west. In the south there is a sudden reversal of this trend close to the carbonatite. As mentioned before pyroxenite is here the characteristic contact rock. To the north, however, the gabbro changes into hornblendite with very little clinopyroxene.

Table 4 lists some chemical analyses of the gabbro. The samples of the layered gabbros were large and contained several light and dark bands.

PERIDOTITES

Peridotitic rocks cover large areas on Stjernøy but within the area mapped in Fig. 2 they are subordinate occurring just as lenses of varying size within the layered gabbro, the gabbro gneiss 1, and the hornblendite.

The peridotites are younger than, and contain inclusions of irregular fragments of the layered gabbro into which they were intruded.

Where peridotites occur in the gabbro gneiss, and especially in the hornblendite, their outer margins are characteristically secondarily altered with much serpentinization of the olivines. When they are of somewhat larger size their central parts usually are unaffected by any secondary alteration.

Within the layered gabbro the contacts between the gabbro and the peridotites are always sharp and well defined.

Within the hornblendite and gabbro gneiss irregular veins of plagioclase hornblendite (and scapolite hornblendite) often penetrate into

Table 4. Chemical analyses, and CIPW molecular norm of four rocks from Table 3 (anal. K. S. Heier), and of hornblende crystal from hornblendite (anal. O. Hageberg in Hysingjord, 1960).

	8	*	5	*	4	*	3	*	Hornbl.	*
SiO_2	43.68	(41.4)	43.21	(40.9)	42.24	(39.6)	34.39	(34.2)	40.28	(38.1)
TiO_2	2.59	(1.8)	4.31	(3.1)	3.92	(2.8)	4.08	(3.0)	3.86	(2.7)
Al_2O_3	15.38	(17.2)	25.73	(28.6)	26.24	(28.9)	7.45	(8.7)	15.48	(17.2)
Fe_2O_3	4.24	(3.0)	5.42	(3.8)	3.72	(2.6)	12.33	(9.2)	3.05	(2.2)
FeO	6.91	(5.5)	7.84	(6.2)	9.10	(7.1)	8.53	(7.1)	7.71	(6.1)
MnO	0.03		0.27	(0.2)	0.20	(0.2)	0.44	(0.4)		
MgO	8,23	(11.7)	4.80	(6.8)	4.79	(6.7)	7.16	(10.7)	13.01	(18.5)
CaO	15.60	(15.8)	3.97	(4.0)	2.81	(2.8)	19.89	(21.2)	14.24	(14.5)
Na_2O	1.69	(3.1)	3.11	(5.7)	3.82	(6.9)	1.21	(2.3)	0.16	(0.3)
K_2O	0.25	(0.3)	0.39	(0.5)	0.68	(0.8)	0.38	(0.5)	0.08	(0.1)
P_2O_5	0.10	(0.1)	0.18	(0.1)	1.04	(0.8)	2.90	(2.4)		
CO_2	0.05		0.06		0.53	(0.7)	0.14	(0.2)	0.22	(0.3)
Cl	n.d.		n.d.		n.d.		n.d.		0.15	
$H_2O^{\div}$	0.27		0.19		0.08		0.16			
H_2O^{+}	0.59		0.36		0.45		0.30		1.25	
Sum	99.61		99.84		99.62		99.36		99.49	
Q	—		6.1		4.8		—		—	
Or	1.5		2.5		4.0		2.5		0.5	
Ab	9.0		28.5		34.5		—		1.5	
An	34.5		20.0		4.5		14.8		42.0	
Ne	3.9		—		—		6.9		—	
Calc.	—		—		1.4		0.6		0.6	
C	—		14.4		19.4		—		—	
Ol	7.8		—		—		—		19.5	
En	—		13.6		13.4		—		2.0	
Fs	—		2.8		6.4		—		2.0	
Di	34.8		—		—		42.8		23.2	
Wo	—		—		—		6.2		—	
Mt	4.5		5.7		3.9		13.8		3.3	
Il	3.6		6.2		5.6		6.0		5.4	
Ap	0.3		0.3		2.0		6.4		—	

* cation percent.

and through the periodites. The contacts are transitional with an increase of hornblende in the peridotites towards the surrounding rocks.

In the unaltered peridotites olivine and clinopyroxene make up 80—90 per cent of the minerals present. The olivine/pyroxene ratio

Table 5. Chemical analysis and molecular norm of central part of peridotite, east side of Raafjell. (anal. K. S. Heier).

	Wt %	Cat %	Norm.
SiO_2	40.43	37.4	Or. 1.5
TiO_2	1.25	0.9	Ab. 4.2
Al_2O_3	5.30	5.8	An. 10.8
Fe_2O_3	3.67	2.5	Ne. 1.0
FeO	7.16	5.5	
MnO	0.14	0.1	Ol. 53.7
MgO	27.40	38.0	Di. 22.8
CaO	8.18	8.1	Il. 1.8
Na_2O	0.69	1.2	Mt. 3.7
K_2O	0.27	0.3	Ap. 0.3
P_2O_5	0.11	0.1	
CO_2	0.71	—	
$H_2O\div$	0.25	—	
H_2O+	3.52		
Sum	99.08		

varies within wide limits but olivine is typically most abundant. Orthopyroxene is present as rimes around the olivine grains. Plagioclase (An 60—70), hornblende, opaques, spinel and biotite are also typically present.

At the hornblendite and gabbro gneiss contacts the olivine is altered to serpentine. Calcite (or another carbonate mineral; magnesite?) is often present close to these contact zones.

A chemical analysis of the central part of a comparatively large peridotite within the hornblendite at Råfjell is given in Table 5. Microscopic examinations showed this sample to be least affected by any secondary alteration (including the peridotites in the layered gabbro). The high water content (3.5 per cent), and also the CO_2 content, indicates, however, that some alteration has taken place, and that the thin section was probably not representative of the analysed sample. The norm in Table 5 is calculated without water, and is very similar to the mode.

HORNBLENDITE

The carbonatite is surrounded on all sides by hornblendite, except in the south where it is in contact with the nepheline syenite, and in the far south east where it borders on the pyroxenite

This hornblendite massif is the only one studied in this report. Hornblendite also occurs at Landvindsfjell and Bukkerbuktsfjell but will not be described here.

Hornblendite is rather common and occurs at several other places on Stjernøy, especially on the west.

The map (Fig. 2) indicates a possible genetic relationship between hornblendite and carbonatite. Most of the hornblendites in other parts of the Seiland province are not visibly connected with carbonatites, and the genetical connection between them is not always clear from the general field evidence. However, where smaller carbonatite bands occur elsewhere they are always connected with hornblendite, or "hornblendization" in such a way as to suggest a genetic relationship.

What is termed hornblendite in the field includes a variety of hornblende-rich rocks of varying structure, grain size, and mineral composition. In all of them hornblende is the dominant mineral ($>75\,\%$ hornblende). However, all gradations between hornblendite and gabbro gneiss occur, and exact contacts are nowhere exposed. Likewise the change from hornblendite into pyroxenite along the southeast carbonatite contact is gradual.

Opaque minerals (magnetite and ilmenite), and apatite are nearly always present; clinopyroxene, biotite, and calcite are common; plagioclase, scapolite, spinel and sphene may be present. When the hornblendite grades into gabbro gneiss the plagioclase content increases rapidly.

The grain size varies within wide limits from fine-grained into pegmatitic varieties. In the latter hornblende crystals may attain 30—40 cm. A chemical analysis of hornblende from the pegmatitic variety, from which the other minerals were carefully separated by the means of a magnet separator, is given in Table 4; pleochroism, x-light yellow, y-brown, z-brown; $2V_x$ is large (80), (data from Hysingjord, 1960).

The hornblende formula calculated on the basis of 22 oxygen atoms (excluding hydrogen-bonded oxygen) is: $(K_{0.02}Na_{0.04}Ca_{2.14})$ $(Mg_{2.79}Fe^{2}_{0.92}Fe^{3}_{0.33}Ti_{0.41}Al_{0.36})$ $(Al_{2.24}Si_{5.67})$ $(OH_{1.20}Cl_{0.04}(F)_{0.76})O_{22}$.

The hornblendite-carbonatite contact is typically sharp. However, although the central parts of the carbonatite are biotite-banded, hornblende is characteristic close to the contact, and calcite veins are common in the hornblendite close to the carbonatite. The smaller carbonatite occurrences elsewhere on Stjernøy are hornblende carbonatites rather than biotite carbonatites.

The hornblendite — gabbro contacts are always transitional. Going towards the gabbro gneiss in the west the contacts are very transitional. Typical hornblendite may "break out" within the gabbro gneiss, and inside massive hornblendite bits of foliated gabbro gneiss may occur.

Leucocratic garnet gneisses are also seen within the hornblendite, and can be traced continuously from where they occur in gabbro gneiss.

As stated earlier peridotites occur in the hornblendites but tend to be altered along the contacts. Veins of hornblendite (or rather hornblende plagioclase veins) are often seen cutting into and through the peridotite inclusions.

CARBONATITE.

The carbonatite covers an area of about 4 km^2. The other occurrences of carbonatite on Stjernøy are small, and have not been studied in his report.

Strand (1952) described the carbonatite. He used the name biotite søvite. Søvite is the local name for the niobium containing carbonatite in the Fen area, S. Norway. The use of this modified local name for the Stjernøy carbonatite should be discontinued. The two rocks are in several respects rather dissimilar. Nb and rare earth minerals which are typical in the Søvite have not been found on Stjernøy. The only thing common to them both is their classification as carbonatites (i.e. a metasedimentary nature is not obvious).

The use of the term carbonatite for the rock on Stjernøy is based on the general usage: "the term carbonatite has been retained for those carbonate-rich rocks that might have originated from a hot carbonatic fluid genetically derived through some magmatic process." (PECORA, 1956, p. 1538) (i.e. where a metasedimentary nature is not obvious).

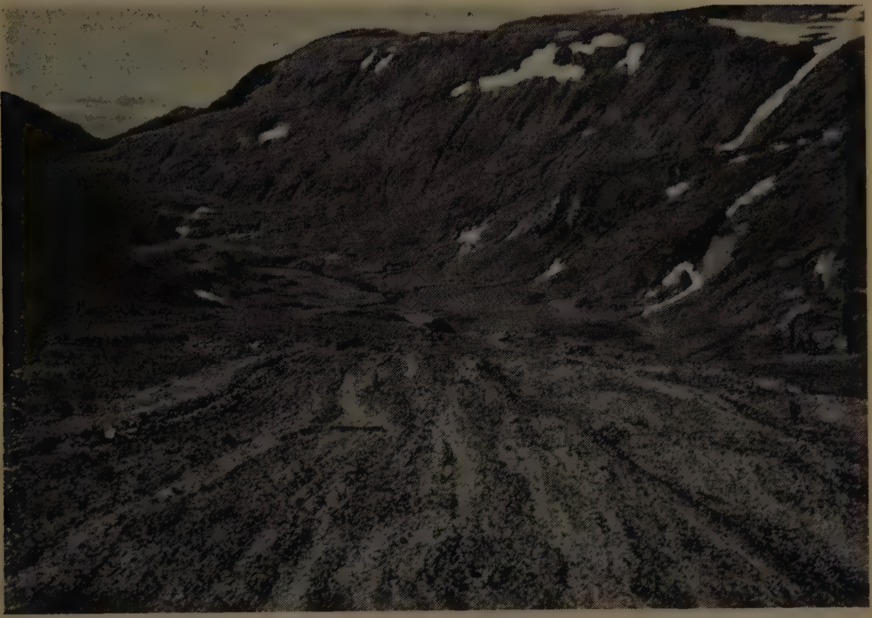


Fig. 6. Biotite bands in carbonatite.

The most remarkable morphological feature of the carbonatite is the separation into biotite rich and calcite rich bands, Fig. 6. This banding accords with a suggestion of a metasedimentary origin but it may also be explained otherwise (p. 150).

Bands of nepheline-albite rocks and pegmatites are very common within the carbonatite, in particular immediately north of the nepheline syenite on Nabberen, Fig. 2.

The carbonatite, even if originally magmatic, today apparently must be classified as a metamorphic rock. Stretching of the biotite bands into boudin-like structures (Fig. 7) is common, and the nepheline pegmatites, which for geological reasons are assumed to be younger than the carbonatite, exhibit similar structures. In places the calcite can be seen to have recrystallized into very coarse grains.

The carbonatite is not foliated. The strike and dip directions indicated on Fig. 2 are the directions of the bands.

Because of the weathering it is difficult to obtain representative hand specimens of the carbonatite. A mineralogical study was done

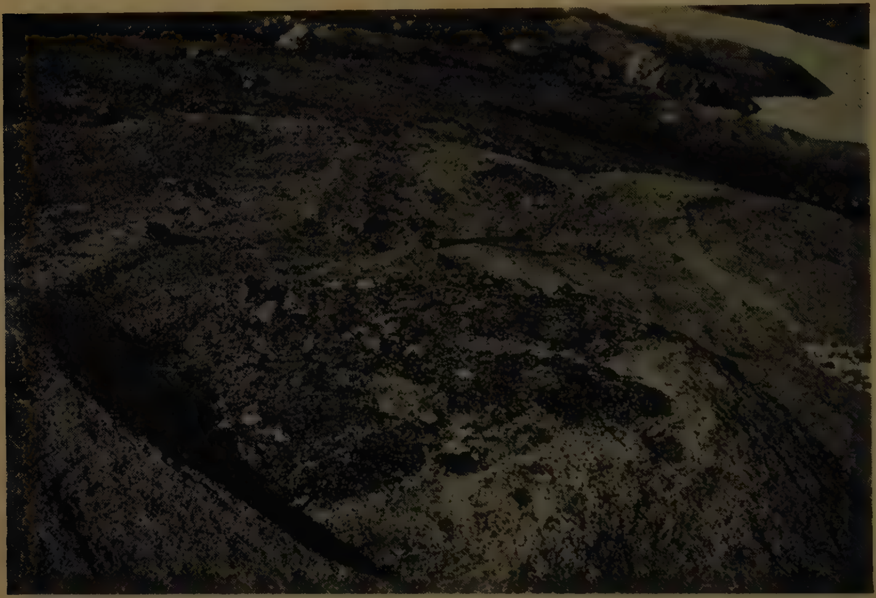


Fig. 7. Boudinage structures in carbonatite.

on material scraped into paper bags from the weathered surfaces. More than 20 different localities from all over the carbonatite were sampled in this way. So far they have not been studied in detail but they appear to be mineralogically uninteresting. None of the Nb and rare earth minerals typically associated with carbonatites elsewhere are found.

The minerals that are found in the carbonatite are:

(a) *Calcite.* A chemical analysis of calcite from the carbonatite is shown in Table 6. Semiquantitative spectrographic Sr and Ba determinations on three metalimestones (marbles) occurring south of Stjernsund (mapped by KRAUSKOPF, 1954) are shown in the same Table. The carbonatite is higher in SrO by a factor of 4—5, and in BaO by a factor of 10. The same trend, though much more pronounced, was shown by Holmes (1959) for other carbonatites.

(b) *Apatite.* This is a very common mineral everywhere within the carbonatite. Its concentration varies from place to place. Chemical

Table 6. Chemical analysis of calcite in:

	Carbonatite Gammevann	Crystaline limestone*		
		Øksfjordneset	Ytre — Lokkerfjord	
CaO	52.26	—	—	—
FeO	1.25	—	—	—
MgO	0.73	0.88	—	—
MnO	0.20	—	—	—
SrO	1.75	0.40	0.40	0.30
BaO	0.11	0.03	0.02	0.01
CO_2	43.67			
Sum	99.97			

(anal. Hysingjord, 1960 — SrO, BaO determined spectrographically.
* Samples supplied by Heier).

analyses and thin section analyses of STRAND (1952) show apatite concentrations to vary between 1 and 10%. It appears that 5% is a good average for the carbonatite as a whole.

The following data were determined for the apatites (HYSINGJORD, 1960):

Cl	0.06%
H_2O	0.00%
CO_2	0.60%
n_0	1.642

The apatite was calculated from the diagrams of Tröger (1956):

Cl-apatite	1%
CO_2-apatite (dahlite)	10%
Oxyapatite	71%
F-apatite	18%

(c) *Eiotite.* Next to calcite biotite is the most common mineral in the carbonatite. Most of it is concentrated in biotite rich bands (Fig. 6) but even the light calcite-dominated bands are fairly high in biotite. In hand specimens the biotite is very dark, and in thin sections the pleochroism varies from light yellow to nut brown. Thin section studies reveal that the biotite is often bent and twisted. Remembering the observed boudin structures this is hardly surprising. The biotites may be several cm large, and show crystal form.

(d) *Hornblende* is especially common in the carbonatite close to the hornblendite contacts. Close to the contact hornblendite lenses are often seen in the carbonatite, and they may also form boudins. In the central part of the carbonatite hornblende is rare or non-existent.

The relative occurrences of hornblende and biotite are considered to be genetically important, and are discussed later.

(e) *Iron ore* is always present in the carbonatite in amounts up to 5%.

(f) *Nepheline* and *albite* are often present. However, in these cases it is difficult to state whether or not they have been introduced later. The map, Fig. 2, shows nepheline gneisses and pegmatites to penetrate into the carbonatite with a maximum frequency close to the nepheline syenite.

SYENITE.

As mentioned before some of the leucocratic gneisses included in the gabbro gneiss 3 are of syenitic composition. In this section another type of syenite is described.

A syenite body occurs within the layered gabbro, Fig. 8. It is elongate in an east-west direction parallel to the banding of the gabbro. It is more than 1 km long and has a maximum width of 150—200 metres. The western end is covered by boulders but towards the east the syenite branches out into the gabbro. Close to the gabbro contact the syenite and gabbro alternate as thin parallel bands parallel to the contact, Fig. 9. In the field this banding is very similar to the characteristic anorthosite-pyroxenite banding of the gabbro (Fig. 3). In some places migmatite-like contacts are seen, Fig. 10.

Alkali feldspar is by far the dominant mineral in this rock. The albite and potassium feldspar occur as separate lamellae and the feldspar is a mesoperthite. In some thin sections independent plagioclase grains outside the alkali feldspar are seen. However, these may well be products of exsolution.

Clinopyroxene is typically next in importance, and occurs in amounts of about 5%.

Opaques and apatite are common accessories. Zircon and biotite

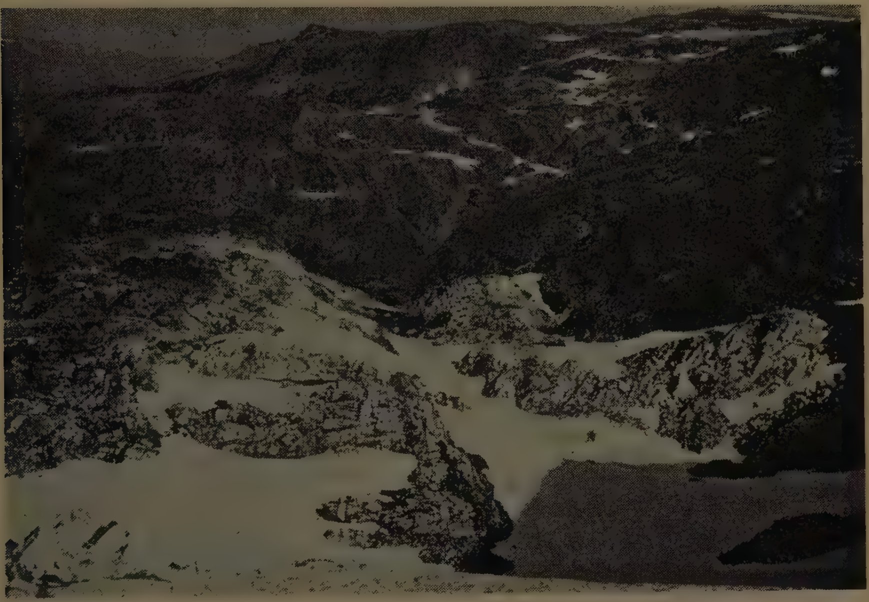


Fig. 8. Syenite in layered gabbro, Lassefjordfjell.

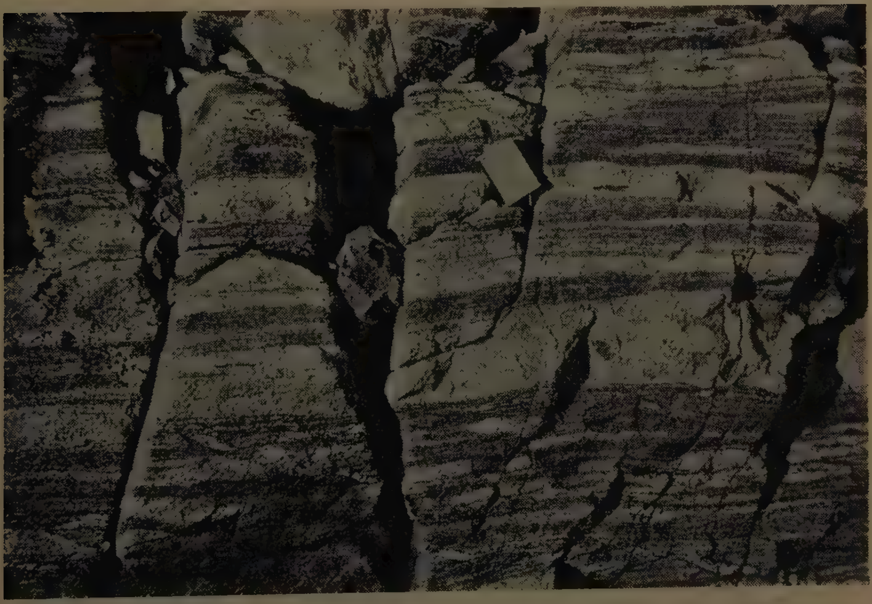


Fig. 9. Alternating syenite and gabbro bands close to gabbro — syenite contact, Lassefjordfjell.

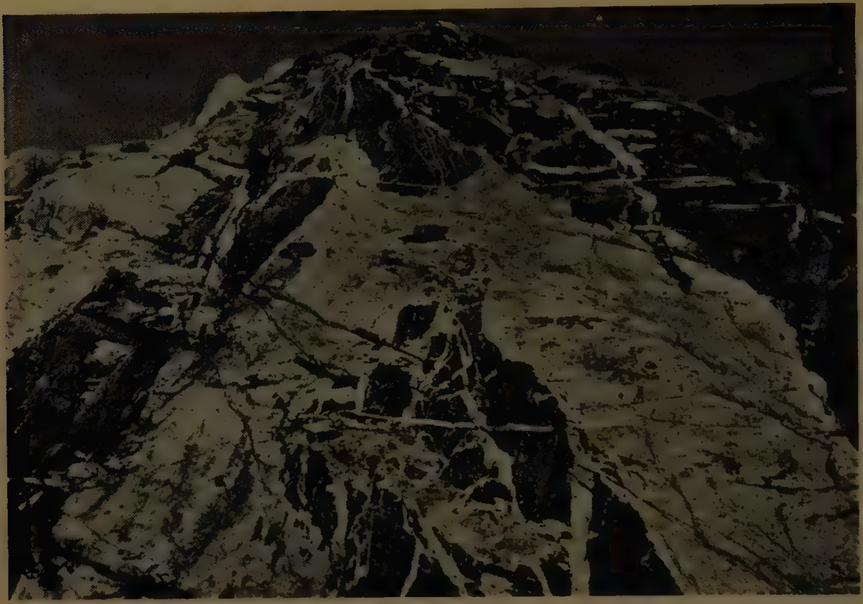


Fig. 10. "Migmatite-like" contact between syenite and layered gabbro, Lassefjordfjell.

Table 7. Chemical analysis, and molecular norm of syenite in layered gabbro, Lassefjordfjell (anal. K. S. Heier).

	Wt. %	Cat. %	Norm.
SiO_2	55.94	51.8	
TiO_2	1.79	1.2	Or. 32.0
Al_2O_3	16.40	17.9	Ab. 43.7
Fe_2O_3	5.61	3.9	An. 3.5
FeO	4.07	3.1	Ne. 4.1
MnO	0.32	0.2	Calc. 0.6
MgO	0.99	1.4	
CaO	3.57	3.5	Wo. 1.6
Na_2O	5.62	10.1	Di. 5.6
K_2O	5.41	6.4	Mt. 5.9
P_2O_5	0.26	0.2	Il. 2.4
CO_2	0.21	0.3	Ap. 0.5
H_2O^+	0.18		
H_2O^+	0.21		
Sum	100.52		

may or may not occur. Calcite may be present in amounts up to 1% but is absent in many thin sections.

A chemical analysis of a several kgm large sample of this syenite is shown in Table 7. Its position on the Nepheline-Kaliophilite-Quartz diagram is shown on Fig. 15.

NEPHELINE SYENITE.

Nepheline-bearing pegmatites are common in the Seiland province including Stjernøy. However, only in one place does a regular nepheline syenitic rock occur. The position of the nepheline syenite is shown on the map Fig. 2. It occurs on Nabberen, between Lillebukt and Storebukt.

The nepheline syenite is in contact with the carbonatite on the east and north. The contact zone of the carbonatite is characterized by nepheline-bearing skarn gneisses, often strongly foliated. Massive nepheline syenite bands also occur in the carbonatite. The map Fig. 2 shows that nepheline gneisses, and nepheline pegmatites penetrate far into the carbonatite towards the north.

Nepheline syenite bands also occur in the silicate gneisses towards the west. They are only slightly foliated, or not foliated at all. They strike northwest and dip steeply to the southwest. Their nepheline content decreases towards the west, and along the shore the light bands are dominantly potassium feldspar (orthoclase) and albite with no nepheline.

The leucocratic bands appear to be young in the gneisses. Brecciation or migmatization of the gneisses as a result of their emplacement are well exposed in shore sections, Fig. 11.

The major nepheline syenite body on Nabberen is lens shaped with the longest axis parallel to the general strike in the area. It has the form of a dike or stock.

Fig. 12 shows the outline of the nepheline syenite. (The map is based on aerial photographs). The syenite extends about 1800 metres in the longest direction with a maximum width of about 250 metres. The exposed area is about 250000 m^2.

Towards the south it pinches out at about 200 m a.s.l., and to the north at about 420 m a.s.l. Its highest point on the mountain plateau is 700 m. (All measurements with pocket barometer).

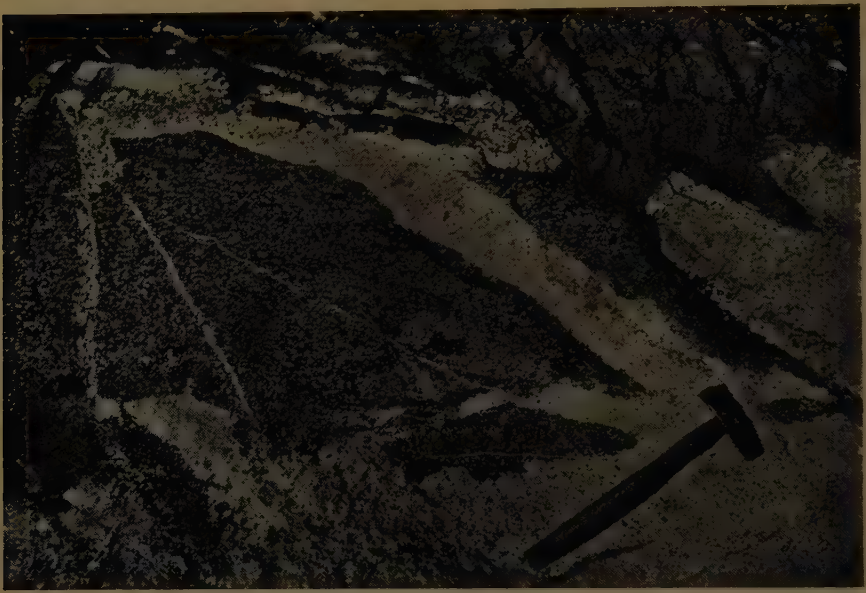


Fig. 11. Migmatitic contacts of leucocratic bands in silicate gneisses, Storebukt.

The nepheline syenite has considerable interest as raw material for the ceramic and related industries. At present, mining possibilities are being examined by Christiania Spigerverk A/S. Because of its economic interest the nepheline syenite was investigated in some detail.

The contacts with the country rocks both to the east and west are generally well defined. An exception is the northeastern contact on the mountain plateau (dashed on the map, Fig. 12). This part of the plateau is covered by boulders (of the nepheline syenite). The immediate country rock is here a nepheline gneiss very similar to the nepheline syenite.

Fig. 13 shows the southern part of the nepheline syenite.

Foliation and some crushing of minerals is seen close to the eastern contact. This tectonic effect disappears towards the west, and the central and western parts of the nepheline syenite are only very slightly foliated or not foliated at all. The foliation directions are indicated on the map Fig. 12. They are always parallel to the contacts.

The nepheline syenite is a homogeneous, medium to coarse grained,

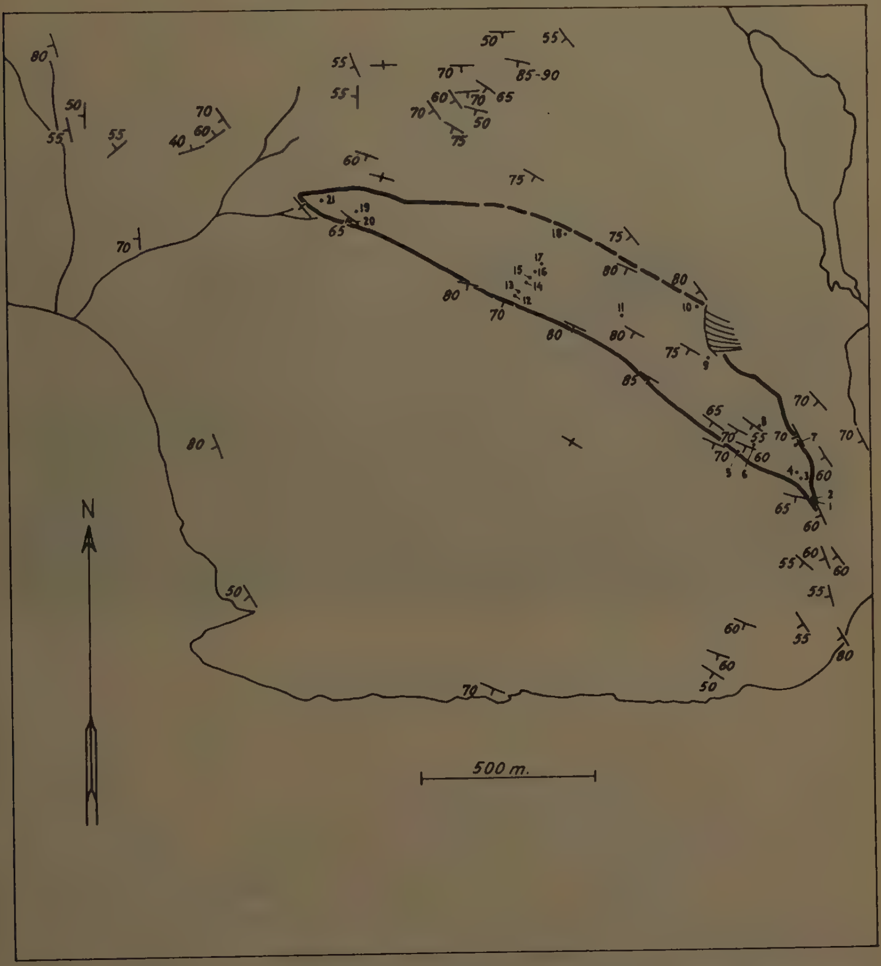


Fig. 12. Outline of the Nepheline Syenite on Nabberen (drawn from aerial photograph). Nos. refer to samples in Tables 8 and 9.

massive rock, in places pegmatitic. The only foreign rocks within it are diabase types of dikes. The field indications are that the dikes are most frequent in the southern part of the intrusion. The strikes and dips of the dikes are variable. The dikes are typically thin (~ 50 cm) but dikes a couple of metres wide are observed. It has so far not been possible to trace any of the dikes into the country rock. However, similar dikes occur in the gneisses west of the nepheline syenite.

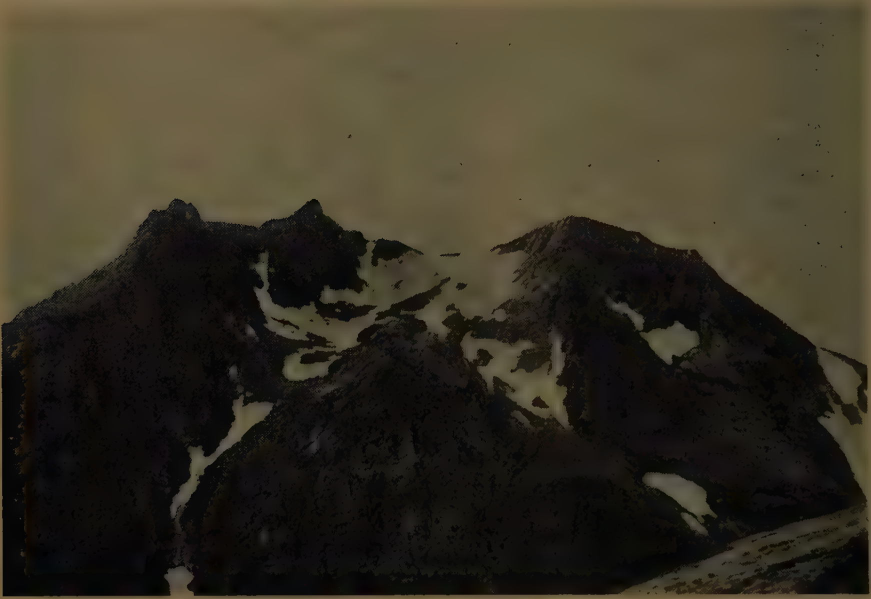


Fig. 13. The southern part of the nepheline syenite, (right ridge) down towards Lillebukt.

Modal analyses (point counted thin sections) of a number of samples from the nepheline syenite are shown in Table 8. Their position in the field is indicated on Fig. 12.

Based on differences in the mode the central part of the nepheline syenite is separated from the portions to the north and south. The latter two are mineralogically very similar. In the following discussion the nepheline syenite in the northern (samples 1—8) and southern (samples 20—21) portions are referred to as Type 1, and the central part (samples 9—19) as Type 2.

The important mineralogical differences between the two types are:

(1) Type 1 is on an average higher in plagioclase (albite as independent grains outside the alkali feldspar perthite).

(2) Nepheline decreases markedly in the extreme north and south ends. However, within the central parts of Type 1 the nepheline content is not very different from that of Type 2.

(3) The K-feldspar content is about the same in both types. It is always an orthoclase perthite (or mesoperthite) which appears opti-

Table 8. Modal analyses (point counted thin sections) of nepheline syenites, nos. refer to Fig. 12.

	1	2	3	4	5	6	6	7
K-feldspar (perthite) . . .	79.4	54.3	59.0	52.0	57.8	55.9	53.0	51.1
Plagioclase	5.9	1.3	10.1	7.9	6.6	2.4	2.9	tr.
Nepheline	—	19.7	18.7	33.5	28.1	32.0	31.6	39.5
(isotropic min. Nosean ?)	—	—	1.1	—	—	—	—	—
Calcite	2.9	2.6	4.8	3.1	1.7	2.1	5.6	6.0
Clinopyrox.	—	—	—	—	—	—	—	tr.
Hornblende	3.1	13.1	tr.	—	—	—	—	—
Biotite	4.3	3.2	5.8	2.5	5.3	4.2	5.6	2.6
Rutile in feldspar	—	—	+	+	+	+	+	+
Titanite	1.6	1.6	—	—	tr.	0.7	—	—
Apatite	0.8	0.3	tr.	tr.	tr.	0.2	tr.	tr.
Opaques	1.3	2.9	0.5	1.0	0.5	2.5	1.3	0.8
Zeolites	—	1.0	—	—	—	—	—	—
Muscovite	—	—	—	tr.	—	—	—	—
Green spinel	—	—	—	—	—	tr.	—	—
K.-feldsp. + Plag. / *Nepheline*		2.8	3.7	1.8	2.3	1.8	1.8	1.3
Average for samples 3 to 8 = 1.8								
Sum sal	85.3	75.3	88.9	93.4	92.5	90.3	87.5	90.6
Averages for amples 3 to 8 = 90.4								

cally monoclinic. The perthitic texture is, however, different, and shows a regular charge from the center of the nepheline syenite towards both the north and south. Figs. 14 a; b; c; illustrate this change. The variation is due to a coarser nature of the albite lamellae from the center and outwards. It was mentioned that independent albite grains are more common in Types 1 than 2. Possibly the albite grains in Type 1 were originally contained in the alkali feldspar as well, and later exsolved.

(4) Calcite is on an average higher in Type 1.

(5) A very important difference is indicated by the distribution of clinopyroxene, amphibole, and biotite.

Clinopyroxene and amphibole are characteristic of Type 2, while biotite is by far the dominant femic silicate mineral of Type 1. The clinopyroxenes and amphiboles are optically diopside-hedenbergites and common hornblendes respectively. They have not the characteris-

Table 8. (contd.)

	8	9	10	11	11	11	12	13
K-feldspar (perthite) . . .	49.2	58.3	54.9	64.9	50.3	50.5	65.2	57.7
Plagioclase	10.2	tr.	—	—	tr.	tr.	—	tr.
Nepheline	30.6	37.2	33.2	30.2	43.4	43.3	32.0	38.8
(isotropic min. Nosean ?)	—	tr.	—	—	—	—	—	—
Calcite	2.2	2.5	4.1	2.0	2.3	2.3	1.6	0.6
Clinopyrox.	—	—	3.8	1.3	1.7	1.7	tr.	2.2
Hornblende	—	1.0	1.0	0.3	0.7	0.7	tr.	tr.
Biotite	1.7	tr.	—	—	—	—	—	tr.
Rutile in feldspar	+	—	—	—	—	—	—	—
Titanite	—	tr.	1.1	0.5	0.7	0.5	tr.	0.5
Apatite	—	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Opaques	0.7	1.0	1.9	0.8	0.9	1.0	1.2	0.2
Zeolites	5.4	—	—	—	—	—	—	—
Muscovite.	tr.	tr.	—	—	—	—	—	—
Green spinel	—	—	—	—	—	—	—	—
$\frac{\textit{K.-feldsp.} + \textit{Plag.}}{\textit{Nepheline}}$	1.9	1.6	1.7	2.1	1.1	1.2	2.0	1.5
Average for samples 9 to 19 = 1.6								
Sum sal.	90.0	95.5	88.1	95.1	93.7	93.8	97.2	96.5
Average for samples 9 to 19 = 94.1								

tic alkali pyroxene and amphibole optics. The silica-poor rather than alkali-rich character of the rocks on Stjernøy is important and must be emphasized.

(6) The titanite vs. rutile-in-K-feldspar relations are interesting. Titanite is characteristically present in Type 2 but tends to be absent from Type 1 (except in the extreme south end, nos. 1, 2, Table 8). What is termed "rutile in feldspar" are thin needle-like inclusions in the potassium feldspar (sagenite structure). The needles are assumed to be rutile but this is not definitely known. A colorimetric TiO_2 determination on nepheline-feldspar concentrates of samples 6 (Type 1) and 11 (Type 2) showed respectively 0.02 and 0.03% TiO_2. Thus the feldspars without the sagenite structure are highest in Ti.

Feldspars crystallizing at high temperatures may take some Ti into solid solution. At lower temperatures such feldspars are unstable and TiO_2 is exsolved. In this connection it should be pointed out that the feldspars showing the sagenite structures also have the coarsest

Table 8. (contd.)

	14	15	16	17	18	19	20	21
K-feldspar (perthite) . . .	58.8	59.4	49.1	58.8	54.1	60.0	55.4	59.0
Plagioclase	0.6	tr.	tr.	tr.	tr.	0.5	4.7	4.6
Nepheline	36.8	32.6	46.6	34.6	36.8	34.9	26.8	27.5
(isotropic min. Nosean ?)	—	—	0.3	—	—	—	—	tr.
Calcite	1.9	2.6	0.3	0.8	2.5	0.9	5.6	5.5
Clinopyrox.	—	—	1.7	2.9	2.1	tr.	—	—
Hornblende	0.4	2.5	0.4	0.2	2.1	1.7	—	—
Biotite	—	1.1	—	—	—	tr.	5.4	1.2
Rutile in feldspar	—	tr.	—	—	—	—	+	+
Titanite	0.2	0.9	0.8	1.2	0.9	0.9	—	—
Apatite	tr.	tr.	0.1	tr.	0.3	tr.	0.2	tr.
Opaques	1.3	0.8	0.7	1.5	1.2	1.0	1.9	2.2
Zeolites	—	—	—	—	—	—	—	—
Muscovite.	—	—	—	—	—	—	—	tr.
Green spinel	—	—	—	—	—	—	tr.	tr.
$\frac{\textit{K.-feldsp. + Plag.}}{\textit{Nepheline}}$	1.6	1.8	1.1	1.7	1.5	1.7	2.2	2.3
Average for samples 20 and 21 = 2.25								
Sum sal	96.2	92.0	95.7	93.4	90.9	95.4	86.9	91.1
Average for samples 20 and 21 = 89.0								

Samples 6 and 11 were taken from two bulk samples, each of about 8 tons, which were blasted out for ore dressing experiments.

albite lamellae (3). Only in feldspars where the exsolution process has been most intense was titanium exsolved.

As the TiO_2 contents in the rocks of Type 1 (without titanite) are, if anything, higher than those of Type 2 (with titanite), (Table 9), it must be assumed that it is the biotite (and opaques) that carry the bulk titanium content in the former. (Only samples 6 and 11 should be compared as samples 1 and 2 are rather special contact rocks.)

No chemical analyses of the alkali feldspars have been made. Their high Na content is, however, evident from the optical studies (Fig. 14 a, b, c) and their crystallization temperature must have been high. This is partly sustained by their monoclinic nature (orthoclases as determined optically). If the feldspars crystallized within the stability field of "orthoclase" or sanidine minimum temperatures of not less than 500° C must be assumed (Laves, 1960).

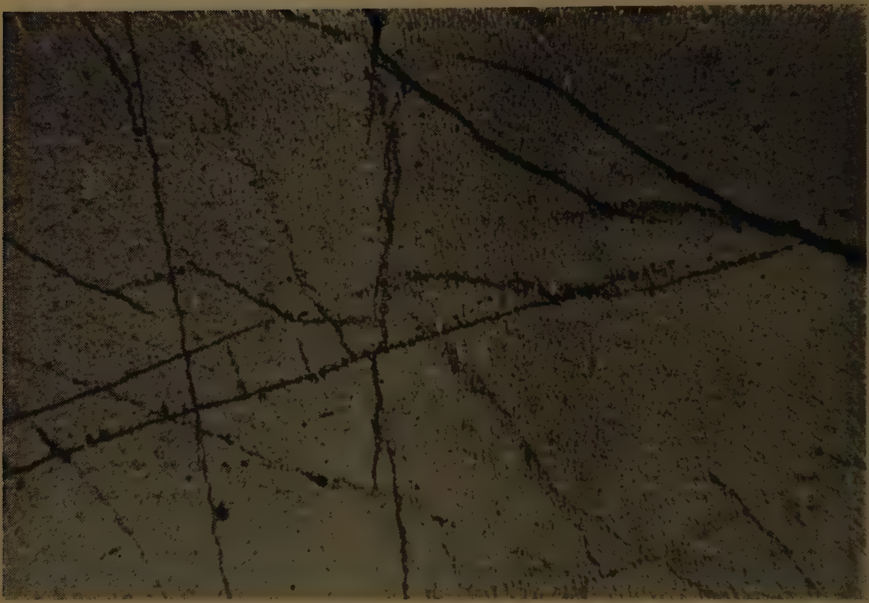


Fig. 14 a, b, c. Variation in the alkali feldspar perthite texture from the center (a), towards the south end (c) of the nepheline syenite (all photographs x 170).

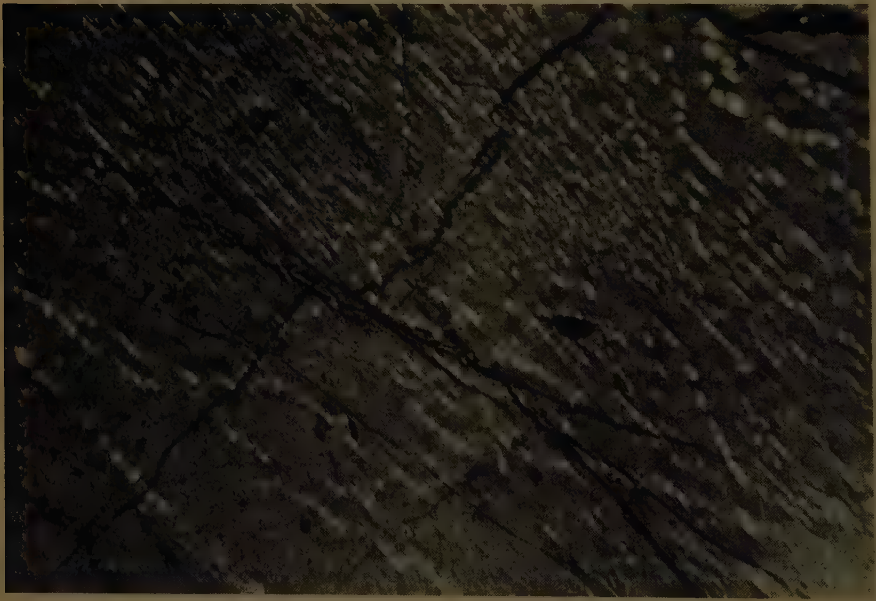


Fig. 14 b.

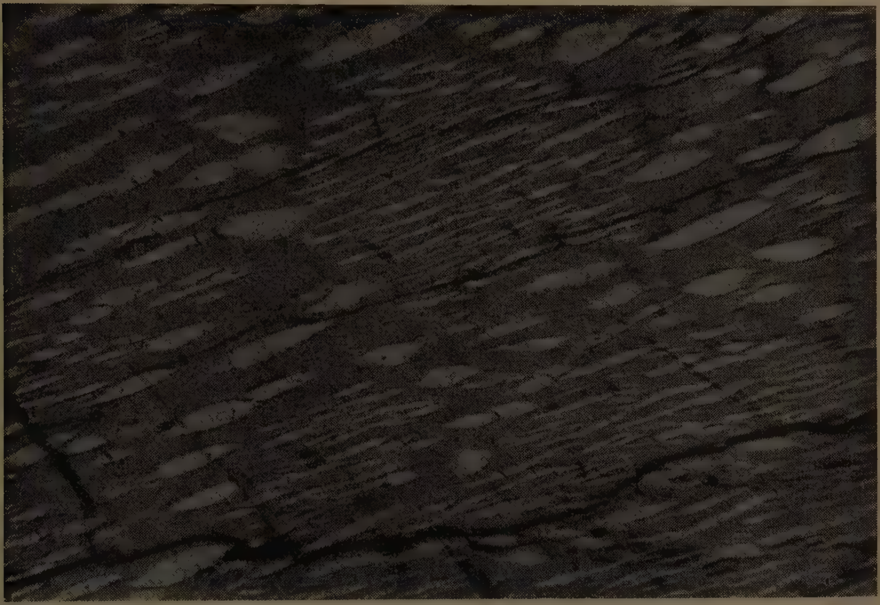


Fig. 14 c.

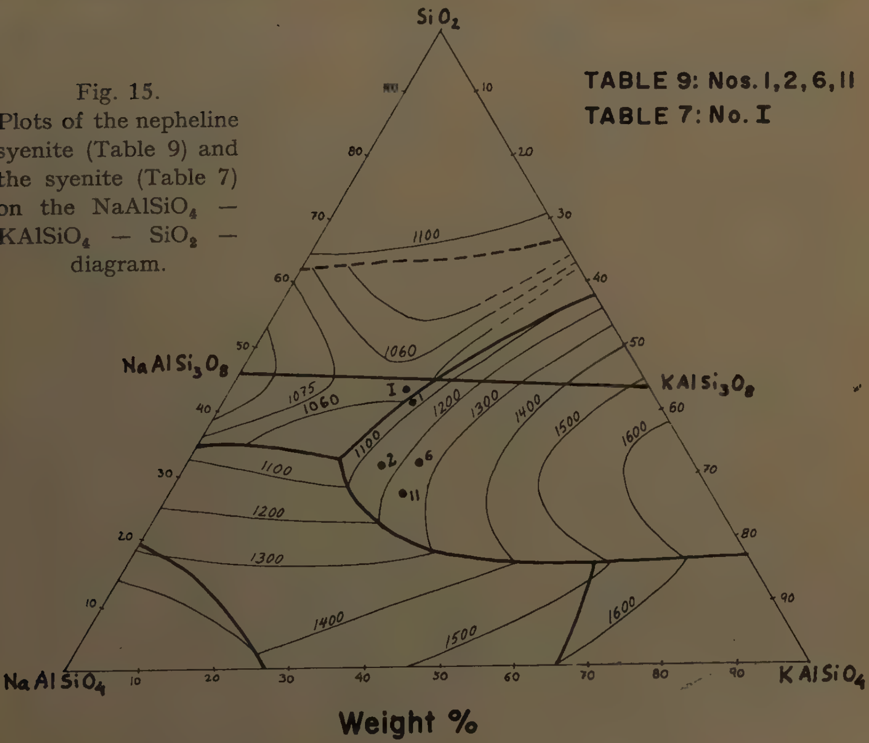


Fig. 15.
Plots of the nepheline syenite (Table 9) and the syenite (Table 7) on the $NaAlSiO_4$ — $KAlSiO_4$ — SiO_2 — diagram.

Table 9. Chemical analyses and molecular norms of four rocks from within the nepheline syenite body (nos. refer to Fig. 12).

	1		2		6		11**	
	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat. %
SiO_2	55.35	50.6	52.69	47.5	52.37	45.6	52.73	46.5
TiO_2	1.57	1.1	1.37	0.9	0.61	0.4	0.51	0.3
Al_2O_3	19.00	20.4	21.88	23.2	23.22	24.3	23.71	24.8
Fe_2O_3	3.00	2.1	2.57	1.7	1.14	0.7	1.89	1.3
FeO	2.67	2.0	2.15	1.6	1.86	1.4	1.04	0.8
MgO	0.59	0.8	0.44	0.6	0.25	0.3	0.24	0.3
MnO	0.07	0.1	0.30	0.2	0.09		0.06	
CaO	4.76	4.7	3.69	3.6	3.11	2.9	2.54	2.4
Na_2O	5.68	10.0	7.47	13.0	6.87	11.8	7.78	13.4
K_2O	5.68	6.6	6.31	7.3	8.30	9.4	8.08	9.2
P_5O_2	0.27	0.2	0.22	0.2	0.09	—	0.05	—
CO_2	1.16	1.4	0.21	0.3	1.88	2.3	0.77	0.9
H_2O+	0.26	—	0.50	—	0.26	—	0.26	—
$H_2O\div$	0.05	—	0.06	—	0.04	—	0.05	—
BaO					*		0.40	
	100.11		99.86		100.09		100.11	
Or.	33.0		36.5		47.0		45.75	
Ab.	37.5		19.6		12.5		6.62	
An.	9.5		2.9		2.5		4.90	
Ne.	7.5		27.4		27.9		36.09	
C.	—		—		2.1		—	
Cc.	2.8		0.6		4.6		1.86	
Ol.	—		—		1.4		—	
Di.	3.2		2.4		—		1.28	
Wo.	0.6		1.6		—		0.24	
Mt.	3.1		2.6		1.2		0.51	
Il.	2.2		1.8		0.8		0.68	
Hm.	—		—		—		0.92	
Ap.	0.5		0.5		0.1		0.11	

* BaO was later determined as 0.47% in this rock.

** Sulfur was later determined as 0.04% in this rock.

Nos. 1 and 2, anal. K. S. Heier.

Nos. 3 and 4, anal. B. Brun.

Chemical analyses of four nepheline syenites taken in succession from the south to the center of the intrusion are listed in Table 9. Their position in the $NaAlSiO_4$-$KAlSiO_4$-SiO_2 diagram is shown on Fig. 15. If the *relative* temperatures read from this diagram are valid the nepheline syenite crystallized from the center and outwards. The position on this diagram of the syenite within the layered gabbro is shown. The crystallization temperature of the syenite is indicated to be lower than for any of the nepheline syenites. Too much significance should not be put even on the relative temperatures indicated by Fig. 15. The close association of nepheline syenite and carbonatite indicates that the nepheline syenite crystallized under a high CO_2,* and probably also H_2O, pressure. This may cause changes in the melting relations.

Possible form of the nepheline syenite towards depth.

Because of the economic interest attached to the nepheline syenite (as raw material for ceramics and related industries) much time was spent in the field with the purpose of estimating the three dimensional form of the intrusion. Based on the field evidence it seems most likely that it has a "stock" like form, the exact form of which it is not possible to predict.

Even though there is some evidence for the nepheline syenite extending towards appreciable depth the chemical and mineralogical composition may change in a vertical direction, as it is observed to do in the horizontal plane. The lowest points of the nepheline syenite in the north and south are respectively 300 and 500 metres below the plateau.

NEPHELINE — ALBITE PEGMATITES.

Nepheline-bearing pegmatites are common within the Seiland province. BARTH (1927) has discussed the different petrographic types in some detail.

On Stjernøy nepheline — albite — biotite — pegmatites are most common. Pegmatites containing mainly albite and some biotite are next in importance. Syenite pegmatites, and also some granite pegmatites are closely associated with the leucocratic bands in the gabbro gneiss. These are, however, rare and always of small size.

---

* "Carbon dioxide under pressure has little or no effect on the rates of melting or crystallization of granite or alkali feldspars, and the melting temperatures are apparently unaffected except insofar as pressure per se raises the temperature of melting." (Wyllie and Tuttle, 1959).

The nepheline-albite-biotite-pegmatites are most common within the carbonatite, and in the layered gabbro east of the carbonatite. Some of them contain calcite (very common in the pegmatites within the carbonatite). Titanite is a common accessory mineral. The pegmatites are not shown on the map (Fig. 2).

One pegmatite occurring west of the north end of the lake north of Lillebukt (Fig. 2) contains; nepheline, albite, sodalite (blue), calcite, garnet (spesartite), apatite, and iron ore. In general, however, the pegmatites are mineralogically uninteresting.

The sizes of the pegmatites vary within wide limits. In the carbonatite between Gammevann and Nabberen the pegmatites may be several metres broad, and extend for hundreds of metres along the strike.

Within the layered gabbro on Lassefjordfjell the pegmatites are characteristically narrow (1/2—2 m). However, they can be traced for hundreds of metres in the strike direction with little variation in width. They are in fact *dikes*. Their strike is normal to the direction of the gabbro bands (about north-south with dips to the west). This direction is parallel to the carbonatite-gabbro contact, and probably represents a direction of opening of parallel fractures in the crust at the time of the carbonatite emplacement.

In spite of their narrowness the pegmatites show up as white stringers on aerial photographs.

The nepheline pegmatites are not foliated, and their minerals show no sign of crushing. However, some of the nepheline pegmatites and nepheline gneisses in the carbonatite show boudinage structures very similar to the biotite bands.

SKARN GNEISSES

Skarn gneisses high in lime silicates occur along the east and the north side of the nepheline syenite, between this rock and the carbonatite. To the south and southeast of the nepheline syenite, carbonatite does not occur, and skarn gneisses are the only rocks between the nepheline syenite and the gabbros to the east (Fig. 2). Calcite bands and lenses are, however, very common within the skarn gneisses all the way down to Lillebukt in the south.

In the south the skarn gneisses are typically dark and heavy while lighter coloured nepheline and calcite rich bands dominate to the north. Both dark and light coloured rocks occur, however, all over the skarn gneiss area. The skarn gneisses have not been mineralogically or petrographically examined in the laboratory. They may, however, very likely prove to be mineralogically interesting.

The foliation directions are always parallel to the nepheline syenite — carbonatite contact (except along a suggested fault line running E—W from Storebukt (Fig. 2)).

GNEISSES BETWEEN THE NEPHELINE SYENITE AND THE SEA TO THE SOUTHWEST.

The gneisses south west of the nepheline syenite are of a banded nature. The dark bands which represent the old complex are pyroxene-hornblende-plagioclase rocks. The light coloured bands are related in time to the intrusion of the nepheline syenite. The nepheline content in these bands decreases towards the west, and the bands are syenitic and monzonitic.

The direction of the banding is parallel to the nepheline syenite contact, and it is a sharp contact between the light and dark coloured bands.

On shore exposures at Storebukt it can be seen how light bands migmatize the dark gneisses. (Fig. 11).

Discussion of petrogenesis

The rocks on Stjernøy are of the type whose genesis is much discussed: layered gabbro, peridotite, hornblendite, carbonatite. Any new bit of information concerning these rocks may be of value, and an attempt will be made therefore, to collect the field and laboratory data into a genetic hypothesis.

First the tentative nature of this synthesis must be stressed. More detailed work both in field and laboratory is necessary, and the results of the work of other geologists now studying the area should be considered. However, as it seems unlikely that I shall continue the work on Stjernøy a complete presentation of my data, information, and ideas about the area is justified.

RELATIVE AGE OF THE ROCKS.

Even though the absolute age, and genesis of the rocks are uncertain, their relative ages seem reasonably clear.

(a) The gabbroic rocks are the oldest rocks within the area. However, the relations between the different gabbros are not as easily ascertained.

The gabbro gneiss 1 with interlayered leucocratic bands is considered as a supracrustal (lavas, tuffs, sediments) component. (It may be noted that gneisses at Øksfjord contain metalimestone (KRAUSKOPF, 1954).) A regional metamorphism under granulite facies conditions separates this rock from the carbonatite-hornblendite emplacement. The texture of the rocks included in this series is typically granulitic. The potassium feldspar is an orthoclase perthite. Hypersthene is a characteristic femic mineral (the alteration of hypersthene into hornblende is a later process which can be demonstrated to be connected with the formation of the hornblendite). The garnets in the leucocratic bands have an almandite/pyrope ratio, Table 10, characteristic of garnets from granulite facies rocks (HOWIE and Subramaniam, 1957). (The garnets are relatively, though not exceptionally, high in Ca, and the lack of ferric iron is remarkable).

Table 10. Chemical analysis and normative composition of garnet from quartz feldspatic gneiss (anal. O. Hagberg in Hysingjord, 1960).

SiO_2	39.17
TiO_2	0.35
Al_2O_3	22.54
Fe_2O_3	0
FeO	26.15
MnO	0.42
MgO	5.84
CaO	6.64
H_2O	0
	101.11

Almandite	57.48 %
Grossularite	18.70 %
Pyrope	22.87 %
Spesartite	0.95 %

The relation between the gabbro gneiss 1 and the other gabbro series in space and time is by no means clear. The layered gabbro and the gabbro gneiss 2 may be lavas. However, the common occurrence of anorthosite bands in the layered gabbro is not compatible with their being lavas. A volcanic interpretation seems to make a subsequent metamorphic differentiation necessary. Indeed both the gabbro gneiss 2 and the layered gabbro are metamorphics. Although textural criteria are not too reliable, the gabbro gneiss 2 shows textures which are generally accepted as metamorphic, and the occurrence of garnet in the layered gabbro suggests a metamorphic origin (= recrystallization) of this rock also.

Barth (1953) implicitly suggested a metamorphic and metasomatic formation of the layered gabbros on Seiland from an originally layered series of supracrustal rocks (sediments and lavas).

With the adjacent supracrustal gabbro gneisses in mind this possibility is not too remote.

However, the layering common to both the Seiland and Stjernøy layered gabbros, and the magmatic Skaergaard layered gabbro on East Greenland (Wager and Deer, 1939) may suggest a similar genesis. The complicating factor in the Seiland province is the subsequent metamorphic and metasomatic reactions (and tectonic movements) overlapping the primary magmatic structures. Nor is it known whether the layering is of the same type in both cases.

(b) *The peridotites* were introduced after the gabbros. Where they occur in the layered gabbro they contain inclusions of the latter. When occurring within the hornblendite (and gabbro gneiss 1) they are secondarily altered (serpentinized). Thus they are considered older than the hornblendites.

(c) *A period of metamorphism* under granulite facies conditions, the effect of which was discussed above, probably preceded the introduction of the younger rocks. The metamorphism possibly also predates the peridotites (and even the layered gabbro — see p. 152).

Because of the superimposing effect of the metamorphic and metasomatic actions related to the intrusion of the younger rocks, the effect of this first metamorphism tends to be concealed. The effect of the younger intrusions was primarily to hydrate the granulite facies rocks with the alteration of hypersthene to hornblende, and considerable related metasomatism.

Intense folding was probably associated with the granulite facies metamorphism. The assumed original horizontal, or near-horizontal, layering of the gabbro gneisses and the layered gabbro was tilted during this folding.

(d) *Faulting* is typically associated with carbonatites and rocks of alkaline affinities. The evidence of faulting prior to, or accompanying the alkaline intrusions on Stjernøy is not immediately obvious but it is assumed on the basis of indirect evidence.

A topographic map of Stjernøy shows two major structural directions which may well be associated with faults. One is about north-south (or N10W), and the other roughly east-west.

The first direction is parallel to the steep and narrow fiords which penetrate into the island from the north. It is suggested that these fiords are along old fault lines, widened during the much later Quaternary ice movements. These faults are traced inland as major drainage directions. The east and west contacts of the carbonatite are parallel to this direction. The east contact coincides with the marked depression running from Lillebukt in the south, and across the island to the steep and narrow Smalfjord in the north. The slightly curved form of this contact, as indicated on the map, suggests a steep west dip of the fault plane. The very sharp carbonatite-hornblendite contact over this distance is indeed consistent with a tectonic border.

A large number of parallel fractures in the north-south direction were evidently opened during this period of faulting, and were later occupied by nepheline-albite-pegmatites.

The east-west faults coincide with another major drainage direction on Stjernøy. This direction is not as topographically conspicious as the north-south direction. However, this has no relation to their relative importance. There is no way to ascertain that the east-west faults were active prior to the intrusion of the alkaline rocks, they certainly were active at some time after these intrusions (see p. 151). Neither is there any direct proof of a vertical displacement associated with the faulting. Possibly only opening of fractures took place. (A vertical displacement may have brought into contact with each other rocks that originated at different depths, i.e., the gabbro gneiss 1 and the layered gabbro).

(e) *The emplacement of the remaining rocks* together with, and after, the faulting, was probably closely connected in time. The following sequence is considered as most consistent with the evidence.

(e1) "*Intrusion*" *of carbonatite.* The nature of the physical state of carbonatite when intruded is unknown and controversial. The use of the name in the present case agrees with the usage recommended by PECORA (1956, p. 1538); "..... the term carbonatite has been retained for those carbonate-rich rocks that might have originated from a hot carbonatic fluid genetically derived through some magmatic process." — and further — "It is clearly implied in the petrologic literature that the carbonatites are a special kind of carbonate-rich rock genetically related to the alkalic rock-forming process and alkalic-magma eruption. The carbonatic fluid, whatever its temperature, concentration, or physical state, is an active, invading agent capable of precipitating minerals."

That a carbonatite magma may exist under normal conditions in the Earth's crust is illustrated by Wyllie and Tuttle (1960). They have demonstrated that the minimum liquidus temperature in the simple CaO-CO_2-H_2O system varies between 685 and 640° C in the pressure interval 27 — 4000 bars.

The carbonatite crystallized to a coarse-grained rock, and whatever its physical state its emplacement was accompanied by gas metasomatism of the surrounding rocks, mainly by H_2O and CO_2.

(e2) *The hornblendite* is assumed to have formed by this type of metasomatic process. In the petrographical part of this paper the growth of hornblende crystals in the gabbro gneisses, and the transitional contacts between hornblendite and gabbro gneiss were described. Likewise, the gradual transition between the layered gabbro, gabbro gneiss 3, and hornblendite (pyroxenite in the south) was described.

However, in some cases hornblendite dikes (or more often hornblende plagioclase dikes) could be seen clearly to cut into the other rocks, illustrating the action of a hornblendite magma. A similar effect may, however, result from metasomatic action of gases. Such gases would preferentially penetrate through cracks and fractures, and strongly alter the wall rocks resulting in dike-like forms.

The existence of hornblendite outside the immediate neighbourhood of the carbonatite is no objection to this mode of formation. It must be remembered that large-scale faulting has taken place, and if the carbonatite originated at some depth the penetration of the gases through other faults is likely.

The transitional gabbros between the carbonatite and the layered gabbro on Lassefjordfjeld give a clear impression of metasomatic processes. The chemical analyses (Table 4) show increasing P_2O_5 and CO_2 concentrations towards the carbonatite, and scapolite may occur both in gabbro and hornblendite.

It is suggested that the areas now occupied by the hornblendite originally were gabbro and gabbro gneiss. In agreement with this is the occurrence of serpentinized peridotites in the hornblendite. The same peridotites when occurring in the layered gabbro are unaltered.

Also pertinent to this discussion and to the composition of the carbonatite magma is the biotite banded nature of the carbonatite (Fig. 6). It is possible that the silicate material was introduced together with the carbonatite. However, it is clear that hornblende bands and lenses are typical close to the hornblendite contacts, while biotite is typical of the central parts. This may be taken to indicate that the biotite bands are metasomatized remnants of the gabbro originally occupying this position. This would necessitate introduction of potassium together with the carbonatite. VON ECKERMANN (1948) suggested that all igneous rocks of the Alnø island alkaline complex originated from predominantly potassic carbonatite magma. Carbonate liquids may exist at temperatures as low as 750° C in systems containing $CaCO_3$ and K_2CO_3 (WYLLIE and TUTTLE, 1960). The low-temperature liquids contain more than 50 weight per cent alkali carbonates. This is much more than the concentration of alkalis normally found in carbonatites. The effect of alkali carbonates on the system investigated by WYLLIE and TUTTLE (1960) has not been investigated.

In summary the field evidence may indicate an intrusion of a potassic carbonatite magma high in both K_2CO_3 and H_2O. The water vapor was the major metasomatizing agent at some distance from the carbonatite altering the gabbro into hornblendite somewhat according to the reaction:

$$\text{pyroxene} + \text{plagioclase} + \text{water} \rightarrow \text{hornblende.}$$

The effect of potassium was restricted to the central parts of the carbonatite where biotite was formed:

$$\text{pyroxene} + \text{plagioclase} + H_2O + K_2CO_3 \rightarrow \text{biotite} + CaCO_3.$$

(e3) *The nepheline syenite* is later than the carbonatite. This is evidenced by the nepheline lenses in the latter, and also by the skarn gneisses. A close connection between the nepheline pegmatites and the nepheline syenite is probable and the frequent occurrence of nepheline pegmatites within the carbonatite indicate the younger age of the syenite. However, the two rocks may well be closely connected in time, and the near association between carbonatite and nepheline syenite must be emphasized. Away from the carbonatite nepheline disappears in the leucocratic bands of the gneisses which are syenitic or monzonitic.

(e3a) The position of the *syenite* within the layered gabbro described on p. 130 is difficult to ascertain. The field evidence gives no clues as to its relation to the nepheline syenite. However, a certain relationship in time would agree with the general picture.

The potassium feldspar of the syenitic rocks were not analysed. However, they are optically shown to be high in albite, and even when assuming a maximum vapor pressure the minimum temperature of crystallization cannot be much less than 700° C.

(e4) It seems safe to assume that a connection exists between the *nepheline — albite — pegmatites* and the nepheline syenite on Nabberen. The nepheline bands and pegmatites in the carbonatite indicate such a connection. They are most frequently met with close to the nepheline syenite. Such a direct connection is more difficult to demonstrate in case of the pegmatites within the layered gabbro. They are explained, however, as fillings of concentric fractures, and may be connected with the nepheline syenite magma at depth.

The existence of a genetic relation between the nepheline syenite and the nepheline — albite — pegmatites needs more elaboration. The nepheline syenite is a nepheline — alkali feldspar rock which contains roughly equal amounts of potassium and sodium. The pegmatites on the other hand are sodium dominated. The only potassium mineral is relatively subordinate amounts of biotite. Very little work has been done on the study of the composition of the aqueous phase in equilibrium with a liquid silicate melt. Some recent work by BURNHAM and TUTTLE (1960) indicate that the aqueous phase coexisting with liquids approaching granitic compositions has higher Na_2O/K_2O ratios (and lower $Al_2O_3/Na_2O+K_2O+SiO_2$ ratios) than the coexisting liquids. Thus the preponderance of sodium relative to potassium in the pegmatites may be in harmony with the experimental data if it is assumed that the pegmatites represent the aqueous phase derived from the syenite liquid.

(f) *Faulting and folding* took place after the carbonatite and nepheline syenite intrusions. The east-west fault which separates the gabbro gneiss 2 from the layered gabbro penetrates the carbonatite immediately north of the nepheline syenite. Variable strike and dip directions are observed in the carbonatite and the skarn gneisess along this fault line.

The uniform north-south foliation and the boudinage structures with biotite and nepheline pegmatite boudins in the carbonatite testify to a late period of folding.

Recrystallization of calcite to large crystals has taken place along certain bands, especially within the skarn gneisses.

The final metamorphism with faulting and folding probably took place at a high level in the crust.

SOME CONCLUDING REMARKS.

The evidence for a period of high-grade metamorphism separating the gabbros from the younger rocks eliminates a comagmatic origin of the Stjernøy rocks. In this connection it must be mentioned that while the evidence of a granulite-facies metamorphism is rather convincing in the case of the gabbro gneiss 1 series, the visible effects of this metamorphism on the layered gabbro (and even gabbro gneiss 2) are more dubious. The metamorphic textures (and minerals) observed in these rocks may have been formed by metamorphic and metasomatic action following the formation of hornblendite.

Assuming the old complex to be represented by the gabbro gneiss 1 series alone, a petrographically interesting comagmatic relationship can be illustrated by the younger rocks, including the layered gabbro.

The original basic magma will be assumed to have been water rich (and rich in CO_2). The gases were kept in the "chamber" during the gabbro crystallization, and a considerable gas pressure was built up as the magma cooled. RINGWOOD (1959) has shown that besides lowering the crystallization temperature, the effect of water is to coordinate preferentially with Al^{3+} in the melt. This preferential Al-OH order will tend to innhibit crystallization of Al^{3+} as $(AlO_4)^{5-}$ groups (in plagioclase). Accordingly Al_2O_3 will tend to accumulate in the residual magma, while CaO, which would have crystallized into anorthite, enters pyroxene (see also MIYASHIRO, 1960, p. 77). Professor Simpson (University of Cape Town) informed me about the close association between gabbro containing clinopyroxene and carbonatites. Orthopyroxene gabbros occurring within the same area cannot be seen to be genetically related to the carbonatites (e.g., gabbro gneiss 1 on Stjernøy). Clinopyroxene is the typical femic mineral in the layered gabbro series.

Yoder studied the influence of water pressure on the equilibrium relations in the system diopside-anorthite (ROY and TUTTLE, 1956). High water pressures enlarge the diopside field. A liquid of composition e.g., $Di_{40}An_{60}$ would under high water pressure first precipitate diopside while under atmospheric conditions anorthite would crystallize first. However, the temperature of crystallization also changes with pressure. Thus the layering of the gabbro cannot be explained by a process of crystallization under rhytmic building up and subsequent release of pressure. If the layered gabbro crystallized from a magma at depth, the mechanism causing the layering may be similar to that postulated by Wager and Deer (1939) for the Skaergaard intrusion.

If the pressure was not released the high concentration of $(OH)^-$ ions together with accumulation of $Al(O,OH)_6^{3+}$ groups will finally induce crystallization of amphibole (RINGWOOD, 1958). In extreme cases the magma may crystallize completely as amphibole taking the Al^{3+} and Na^{1+} (normally crystallizing as albite) and the Ca^{2+} and Al^{3+} (normally crystallizing as anorthite) into solid solution.

A ready inference would be that also hornblendite was formed in this way. However, the evidence for a metasomatic formation of hornblendite is rather convincing. Besides by this type of crystallization a sodium-rich amphibole should be expected which is not the case in the hornblendite. Again it should be pointed out that the Seiland province rocks are characteristically silica deficient but not exceptionally alkaline.

The accumulation of Al and alkalis in the magma during crystallization lead to a syenitic rest magma instead of granitic rest magma of normal fractionation processes. At this stage the built-up gas pressure eventually ruptured the surface, causing large-scale faulting. Intrusion and subsequent crystallization of the gas-rich magma to carbonatite and nepheline syenite together with gases metasomatizing the surrounding gabbro, characterize the concluding stages of the magmatic activity on Stjernøy.

Economic geology

The gabbro dominated Seiland province seems to be barren of economic base metal deposits often associated with basic rocks. Some iron mineralization occurs but are unimportant economically. However, two rock types are of economic interest on Stjernøy.

(1) The nepheline syenite on Nabberen. Mining of this rock as rawmaterial in the ceramics industry has been started by Christiania Spigerverk A/S. Enormous quantities are apparently available. The mining is planned as underground mining with a transport tunnel from west of Lillebukt The ore gives a pure nepheline-feldspar concentrate upon crushing and magnetic separation. This process will be carried out at Lillebukt wherefrom the ready-made concentrate will be shipped.

(2) So far no economic exploitation of the carbonatite is planned. However, the rock, which is high in $CaCO_3$, K_2O (in biotite which easily yields under weathering), and P_2O_5 offers possibilities as fertilizer. The excellent soil given by this rock is demonstrated by the thick vegetation growing on it.

The best way to mine it would probably be to quarry the weathered rock around Lake Gammevann for example. The economic barrier is probably a too expensive transport to the market.

The carbonatite itself is too low in apatite to be promising as a phosphate ore. However, weathering agencies may have caused secondary enrichment of apatite, and such deposits of apatite may be of economic use.

LITERATURE.

BARTH, T. F. W. (1927) Die Pegmatitgänge der kaledonischen Intrusivgesteine im Seiland — Gebiete. Norsk Vid. — Akad. Skr. I, 1927, no. 8.

— (1953) The layered gabbro series at Seiland, Northern Norway. Norg. Geol. Unders. Årbok, 1952, *184*, 191—200.

BURNHAM, C. W. and TUTTLE, O. F. (1960) Composition of the magmatic vapor phase. Geol. Soc. America. Progr. ann. meeting (abstract).

ECKERMANN, H. von (1948) The alkaline district of Alnö island. Sverig. Geol. Unders., Ser. Ca, no. 36.

HOLMES, A. (1958) Spitskop carbonatite, Eastern Transvaal. Geol. Soc. America, Bull., *69*, 1525—1526.

HOWIE, R. A. and SUBRAMANIAM, A. P. (1957) The paragenesis of garnet in charnockite, enderbite, and related granulites. Min. Mag. *31*, 565—586.

HYSINGJORD, J. (1960) En petrografisk beskrivelse av metamorfe og eruptive bergarter på Stjernøy, Vest-Finnmark. Hovedfagsoppgave i mineralogi og petrografi, Universitetet i Oslo.

KRAUSKOPF, K. B. (1954) Igneous and metamorphic rocks of the Øksfjord area, West-Finnmark. Norg. Geol. Unders. Årbok, 1953, *188*, 29—50.

LAVES, F. (1960) Al/Si-Verteilungen, Phasen — Transformationen und Namen der Alkalifeldspate. Z. Kristallogr., *113*, 265—296.

MIYASHIRO, A. (1960) Thermodynamics of reactions of rock-forming minerals with silica. I. Japanese Jour. Geol. Gelgr. *31*, 71—78.

NEUMANN, H. (1960) Apparent ages of Norwegian minerals and rocks. Norsk Geol. Tidsskr., *40*, 173—191.

OOSTEROM, M. G. (1954) En hornblende rik sone i Seiland-peridotitten. Norg. Geol. Unders. Årbok, 1953; *188*, 51—53.

— (1956) Some notes on the Lille Kufjord layered gabbro, Seiland, Finnmark, Northern Norway. Norg. Geol. Unders. Årbok, 1955, *195*, 73—87.

PECORA, W. T. (1956) Carbonatites: A review. Geol. Soc. America, Bull., *67*, 1537—1556.

PETTERSEN, K. (1874a) Om gabbroen paa Seiland, Vest Finnmark. Geol. För. Forh., *2*, 125—126.

— (1874b) Om forekomst av Elaeolith i Vest-Finnmarken. Geol. För. Forh. *2*, 220.

— (1883) Bidrag til de norske kystströgs geologi III. Archiv. Mat. og Naturv., *8*.

RINGWOOD, A. E. (1959) Genesis of the basalt — trachyte association. Beitr. Min. Petr., *6*, 346—351.

ROY, R. and TUTTLE, O. F. (1956) Investigations under hydrothermal conditions. Physics and Chemistry of the Earth, *1*, 138—180.

STRAND, T. (1950) Forvitring og nedising. Norsk Geol. Tidsskr., *28*, 237—239.

— (1952) Biotittsövitt på Stjernøy, Vest Finnmark. Norg. Geol. Unders., Årbok, 1951, *183*, 10—21.

UNDÅS, (1938) Kvartärstudier i Vest-Finnmark og Vesterålen. Norsk Geol. Tidsskr., *18*, 81—217.

WAGER, L. R. and DEER, W. A. (1939) Geological investigations in East Greenland; Part III; The petrology of the Skaergaard intrusion, Kangerdlugssuag. Medd. om Grönland, *105*, no 4, 1—352.

WYLLIE, P. J. and TUTTLE, O. F. (1959) Effect of carbon dioxide on the melting of granite and feldspars. Am. Jour. Sci., *257*, 648—655.

— (1960) The system CaO-CO_2-H_2O and the origin of carbonatite. Jour. Petrology, *1*, 1—46.

Manuscript received February 27, 1961

Printed December 1961.

THE ANORTHOSITIC COMPLEX OF HÅLAND–HELLEREN

By

JEAN MICHOT, Jr
(Université de Liège)

Introduction.

This paper is an abstract of a more comprehensive work published in the mémoires de l'Académie royale de Belgique (J. Michot, 1961). It records information obtained from geological researches carried out in the Egersund area (southwestern Norway) from 1952 to 1958 during six successive summers, each of five or six weeks. The material collected during the field work was studied at the University of Liège (Belgium) in the department of Geology, Petrology and Geochemistry directed by Prof. P. Michot.

The author wishes to acknowledge the help and interest of Prof. T. F. W. Barth of the Geological Museum, University of Oslo, in whose department he had the opportunity to spend a few months in 1953–1954. He also wishes to thank Prof. N.-H. Kolderup who kindly introduced him to the Bergen area. Thanks are also due to Director R. Brun and to I. Dybdahl, S. A. Titania in Sokndal, for providing working facilities during his stays in the Egersund District.

The object of this work is to present briefly the geology and petrology of the anorthosito-leuconoritic complex of Håland–Helleren and thus to contribute to the knowledge of the phenomena operating in the deep zones of the earth's crust. It will be shown that the basic anatexis, locally mentioned in the Egersund area (P. Michot 1955b), is in fact widespread in this area. Indeed, it is a regional process resulting in the separation of a liquid phase, the anatectic leuconorite, from a residual solid phase the para-anatectic anorthosite. Using a term proposed by J. J. Sederholm, the author will call this process basic palingenesis (J. Michot, 1960).

General Geology.

The Håland—Helleren anorthosito-leuconoritic complex belongs to the basic igneous province of southern Rogaland (C. F. Kolderup, 1896, 1914; T. F. W. Barth, 1936, 1945). It is situated between Egersund and Sokndal with its northeastern end extending northward in the direction of Helleland (map, Plate 1).

In the north, between the sea and Helleren, it borders on the norito-granitic band, further east on the anorthositic body of Egersund—Ogna, and in the northeast and the east on the Bjerkreim—Sokndal body. In the south and the west it disappears under the sea.

This complex consists of three main geological units which are, from oldest to youngest:

(I) the Håland anorthosito-leuconoritic body;
(II) the Åseheia and Augendal intrusives;
(III) the Åmdal—Helleren—Rødland anorthosito-leuconoritic body.

These units are cut by dykes of norite, monzonorite, monzonite, and granitic pegmatite trending mostly north-south, and by doleritic dykes trending approximately west-east.

The mineral facies.

The rocks surrounding and constituting the area, whether of a sedimentary or eruptive origin, belong to the granulite facies or more accurately to the facies named mangeritic by P. Michot, and which is typical of the deep catazone (P. Michot, 1948).

This facies is characterized by the instability of muscovite and partially of biotite, and by the stability of associations containing plagioclase (andesine or sodic labradorite), potash feldspar, hypersthene, diopside and quartz. In rocks sufficiently rich in alkalis, the association includes a special feldspar: the mesoperthite.

The geological processes which have determined the formation of this province have developed, or have come to an end, within the deep catazone.

I. The Håland anorthosito-leuconoritic body.

The lithological composition of the body is fairly simple: light and dark leuconorites and, in places, norites as well as anorthosites. The types of association and the structural relations differ, however from one place to another. Two parts are to be considered:

1. the northern part covering the northern and the middle part of the body, characterized by an almost regular gneissic structure or a distinct irregular banding trending approximately west-eastward;
2. the southern part made up of folded gneisses of a more intricate arrangement.

The two parts gradually pass from one to the other by mutual interpenetration: lenses a few meters long typical of the southern part are embedded in the rocks of the northern one.

1. The northern part.

The northern part mainly consists of two lithofacies both representing the extreme members of a series of associations in which the anorthositic and leuconoritic rocks are present in variable proportions. The first lithofacies — gneissic leuconorite — is to be found on the northern border of the Håland body, the second one — anorthosite with pseudo-inclusions of leuconorite — is situated in its middle part.

The different types of association of anorthositic and leuconoritic rocks are connected with the west-eastward gneissic structure which is characteristic of this area (fig. 1). One type can be seen in an anorthositic rock containing regularly shaped lenticular leuconoritic patches ranging in length from 0.10 to 0.80 m, and sometimes even 1 m (anorthosite with pseudo-inclusions of leuconorite).

These patches show gneissic structure parallel to their major axis which itself remains constantly parallel to the structure of the encasing

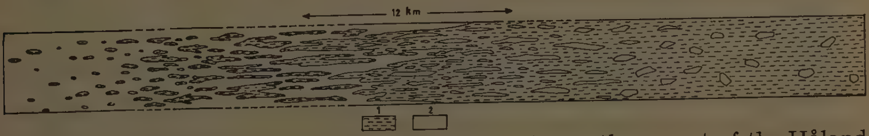


Fig. 1. Anorthosite-leuconorite association in the northern part of the Håland body. 1: leuconorite; 2: anorthosite.

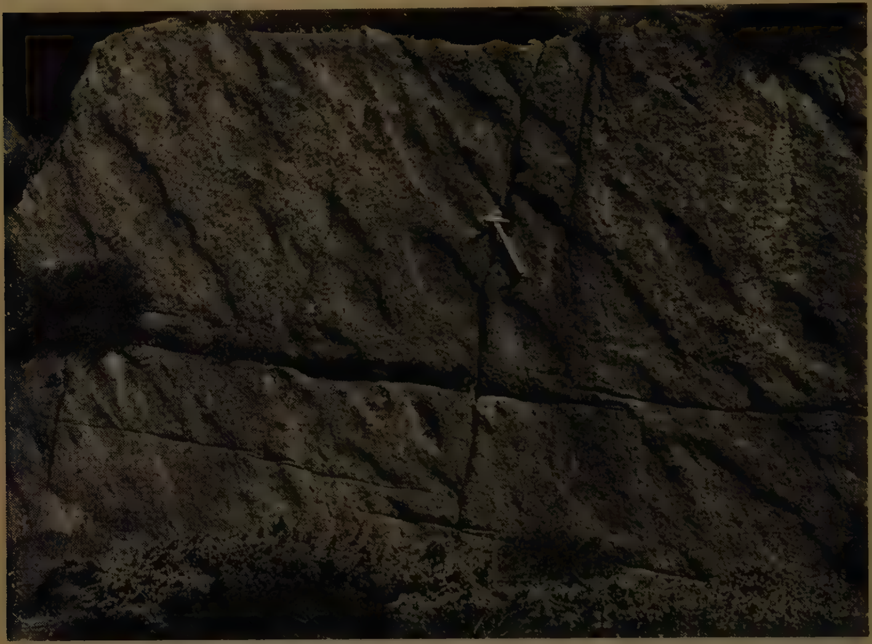


Fig. 2. Anorthosite with pseudo-inclusions of leuconorite.

rock. They sometimes extend into the anorthositic rock through a number of tiny stringers of dark minerals also elongated in the same direction. Therefore they do not present the aspect of true xenoliths (fig. 2).

Along the strike this type of association gradually changes into a succession of lenticular anorthositic beds, 0.30 to 1 m thick and several meters long, alternating with gneissic leuconoritic beds more irregular in shape but of similar thickness. The latter contain locally smaller anorthositic lenses which are sometimes 1 m long. The gneissic structure of the leuconorite follows the borders of the anorthositic beds or lenses and their irregularities and crosses some of the fractures.

Eventually this type changes into gneissic leuconorite with a few lenticular anorthositic inclusions. In places, e.g. on the Kaknuden and on the northern shore of the Kudlandsvd, the inclusions become so numerous as to form a real igneous breccia (basic agmatite). Inclusions of different shape and composition are to be found: massive anorthositic rocks, anorthositic and leuconoritic gneisses. All these xenoliths have sharp angular or sometimes rounded outlines which cut their internal structure. Frequently the structural planes of the xenoliths are set at random in the gneissic leuconorite, and show no relation to the gneissic structure of the encasing rock. Their volume may reach 30—50 cubic meters.

In another place (Hesnes) the gneissic leuconorite, the same which cements the breccia, intrudes into the surrounding rocks of the norito-granitic band.

Petrographical study.

Anorthosite occuring as (1) rock masses embedding the leuconoritic patches (2) beds or lenses alternating with the gneissic leuconorite (3) xenoliths in the igneous breccia, is made up of plagioclase in the composition range An 40 to An 45.

This plagioclase exhibits isometrical or rarely elongate grains with very irregular and sometimes strongly indented outlines. All grains are twinned according to the albite and pericline laws the twins being usually poorly developed and very rarely straight. They very often include thin antiperthitic stringers (0.01 to 0.08 mm in diameter, 0.1 to 0.2 mm long) extending in three directions but sometimes limited to the border of the twins (J. Michot, 1958). Moreover they contain very thin needles of rutile covered with a brownish film, as well as thin dust of pyroxenes and black ore. They are frequently slightly deformed with curved twins exhibiting undulatory extinctions (fig. 3).

These anorthositic rocks have a texture that corresponds neither to the texture of an igneous rock in which the plagioclase forms well developed plates or laths, nor to the texture of a gneiss in which the minerals are orientated more or less parallel to one another either geometrically or optically. Hence neither the term anorthosite which is often given to the plagioclasic rocks of igneous origin nor the term anorthositic gneiss seems to fit them. The term granofels as proposed by R. Goldsmith (R. Goldsmith, 1959) would seem to be appropriate if the anorthositic qualification is added (J. Michot, 1959).

The leuconorite consists of plagioclase of a first type, identical to that of the anorthositic granofels, disseminated in layers of a second type of plagioclase in association with hypersthene of a

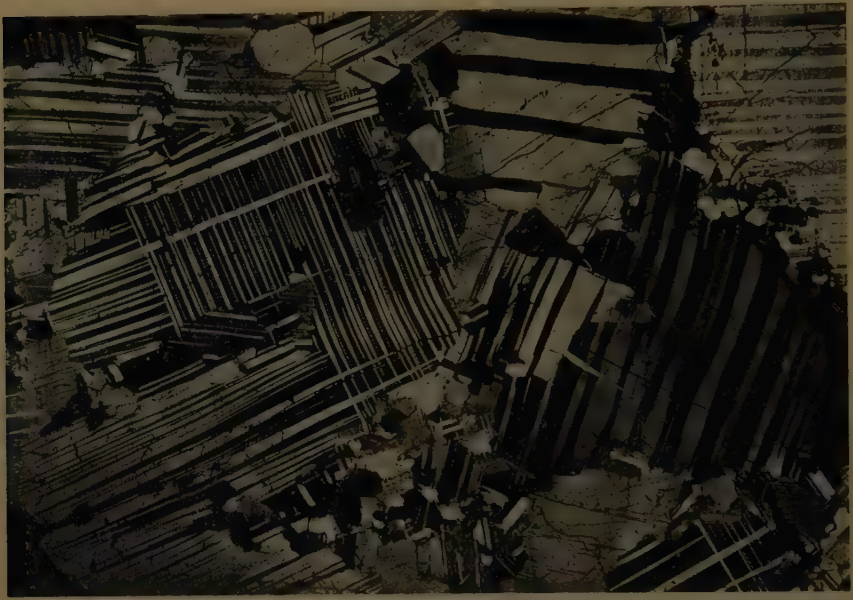


Fig. 3. Anorthositic granofels—Plagioclase of the first type. Nicols +; 11x.

poecilitique interstitielle texture (P. Michot, 1939b), a little diopside and black ore. The second type of plagioclase appears in small grains with regular outlines and clear extinctions, non-antiperthitic and without any dust; the composition varies between 40 and 45% An.

The microscopical study clearly indicates that the plagioclase of the first type is older than the plagioclase of the second type and the ferromagnesian minerals. It also is previous to the gneissic structure which characterizes the leuconorite.

Conclusion: the basic anatexis.

The anorthosite-leuconorite association is characterized by mutual reciprocal relations of these two rocks.

This association may be explained as the result of a differential migration of minerals mainly ferromagnesians and plagioclase of the second type, leaving behind a residue of plagioclase of the first type.

Probably a differential melting, i.e. anatexis, has occured in a pristine rock containing such minerals. Indeed, at the moment of the melting, a eutectoid leuconoritic melt is generated leaving a solid residue composed of the mineral, the quantity of which in the rock is superior to that fixed by the eutectic point, here the plagioclase.

The different types of anorthosite-leuconorite association, are formed by a mechanical differentiation under the influence of shear. The leuconoritic liquid generated by the melting was squeezed out, carrying with it a great number of deformed residual plagioclases (plagioclase of the first type), forming homogeneous masses, sometimes intruding into the country rocks (Hesnes), and sometimes embedding inclusions of residual anorthosite.

Anorthositic granofels left behind during this process in a nearly pure state but locally containing remains of the anatectic liquid (anorthosite with pseudo-inclusions of leuconorite) is the nearest approach to a solid residue.

Anorthositic granofels represents an agglomerate of the residual plagioclases (*para-anatectic anorthosite; P. Michot 1956*). *Leuconorite represents the concentration of the interstitial anatectic liquid* (*anatectic leuconorite*).

2. THE SOUTHERN PART.

This part of the Håland body is composed of anorthositic and leuconoritic rocks in various associations. They often appear as a series of alternating beds 0.30—0.40 m to 1 m thick, made up of anorthositic and leuconoritic granofels, leuconoritic, sometimes noritic, gneisses with small lenses of granulated hypersthene, and gneissic leuconorite; they also form layers 0.05 to 0.20 m or even stripes 0.01 to 0.02 m thick resulting in finely zoned or banded rock types.

On a large scale, the structure of the southern part shows wide undulations. On a smaller scale, the gneissic structure, characterized by the attitude of the ferromagnesian minerals and the small anorthositic lenses parallel to the general trend of this area, is in some places fairly isoclinaly folded with almost horizontal axial planes.

The relation of the gneissic leuconorite to the anorthosito-leuconoritic beds and layers merits special mention (fig. 4).

Locally rather thick beds of leuconoritic granofels contain one or two layers of anorthositic granofels. Along the strike the leuconoritic granofels slowly grades into a gneissic leuconorite whereas the

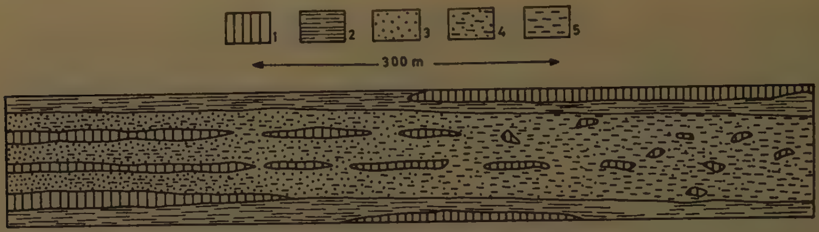


Fig. 4. Relation of the anorthosito-leuconoritic beds and layers to the gneissic leuconorite in the southern part of the Håland body.
1: anorthositic granofels; 2: leuconoritic gneiss; 3: leuconoritic granofels; 4: leuconoritic granofels and gneissic leuconorite; 5: gneissic leuconorite.

intercalated anorthositic layers thin away, break up, and are finally replaced by a series of lenses the length of which varies from 0.50 m to 4 or 5 meters. Further on, the gneissic leuconorite forms a well individualized bed in which small anorthositic inclusions are scattered about.

In the neighbourhood of Mong, a bed of gneissic leuconorite 1 m thick occurs in a series of beds of anorthositic granofels and leuconoritic gneisses. At a special point, the gneissic leuconorite widens, grows thicker and thicker, and intrudes into the surrounding beds. Southeast of Odden, a similar leuconorite constitutes the cement of a breccia which extends over several hundred square meters, and whose inclusions, scattered at random, represent blocks of the gneissic rocks typical of the southern part.

These observations show that the gneissic leuconorite which often forms stiff concordant beds in places behaved as an extremely movable magmatic unit elsewhere.

Petrographical study.

The anorthositic rocks are composed of the same type of plagioclase as that which forms the anorthositic granofels of the northern part.

The leuconoritic rocks are of two types. (1) The gneissic leuconorite is equivalent to the leuconorite previously described and characterized by hypersthene with poecilitique interstitielle texture. (2) Leuconoritic granofels and leuconoritic gneisses carry plagioclase of the first type associated with hypersthene, sometimes diopside, of a prismatic habit. This habit is often observed in the norito-granitic band whose origin is connected to a metamorphic process (P. Michot, 1955a).

Consequently, leuconorites of type 2, forming the major part of the leuconoritic rocks of the southern part, could be of the same origin.

Conclusion.

The transitions observed between the leuconoritic granofels and the gneissic leuconorite, and the fact that the original arrangement of the anorthositic beds or lenses are often preserved, indicate an incipient mobilization of a solid unit towards a magmatic moving one. This mobilization is the expression of an anatectic phenomenon which occurs in the southern part in its formal stage.

The nature of the preexisting rock complex is indicated by observations in the region of Augendal: by the abundance of the beds of anorthositic granofels, by the intercalations of leuconoritic granofels as well as by the presence of small lenses of granulated hypersthene, the rocks become identical with the gneissic border of the anorthositic body of Egersund—Ogna (P. Michot, 1939a).

GENERAL CONCLUSION.

The Håland body is divided into two parts corresponding to two phases of the anatectic phenomenon. The southern part reveals an incipient anatexis whereas, in the northern part, the anatexis has developed on a larger scale, resulting in the production of gneissic anatectic leuconorite and, simultaneously, of para-anatectic anorthosite.

The rocks of the two parts are of a similar nature; this seems to indicate that the Håland body initially formed a homogeneous unit similar to the anorthosito-leuconoritic gneissic border of the Egersund—Ogna body. Consequently it is inferred that the Egersund—Ogna body, the oldest unit of Rogaland, once extended southward to the sea in the direction of Sokndal. Subsequently, south of Egersund, anatectic phenomena interfered and created new types of rock.

II. The Åseheia and Augendal intrusives.

The Åseheia and Augendal intrusives are dark leuconorites, constituting the second main unit within the basic complex studied here. They appear in two places, the first to the northwest, the second to the southeast of the large protuberancy of the third unit, the Åmdal—Helleren—Rødland body, in the west part of this complex.

The dark leuconorites represent typical magmatic intrusions clearly posterior to the anorthosito-leuconoritic body of Håland. They cut across the gneissic or layered structure of this body and contain xenoliths of it and of other rocks very similar to those of the norito-granitic band.

Locally, the dark leuconorites pass into the rocks of the Håland body which they slightly assimilate. This assimilation seems to be responsible for the existence of two main types of leuconorites: a fine-grained and a medium-grained type. The phenomenon is obvious in the surroundings of the gneissic leuconoritic xenoliths which lie within the fine-grained leuconorite; the rock in between is always made up by a medium-grained leuconorite.

Petrographical study.

The dark leuconorites have the same mineralogical composition and the same texture as has the leuconoritic anatectic rock of the Håland body. The minerals that form the association including plagioclase of the second type, pyroxenes (hypersthene and diopside) and black ore, constitute a more important part of the rocks. In this association the hypersthene has a well developed poecilitique interstitielle texture, especially in the fine-grained leuconorite (fig. 5).

Conclusion.

The Åseheia and Augendal magmatic unit is the result of a melting similar to that which generated the anatectic rocks of the Håland body. But in this case the quantity of the molten part was more important, and the residual plagioclases less important quantitatively. The magma, whose mobility was probably rather great, became more completely separated from the residual anorthositic part and was able to intrude the Håland body.

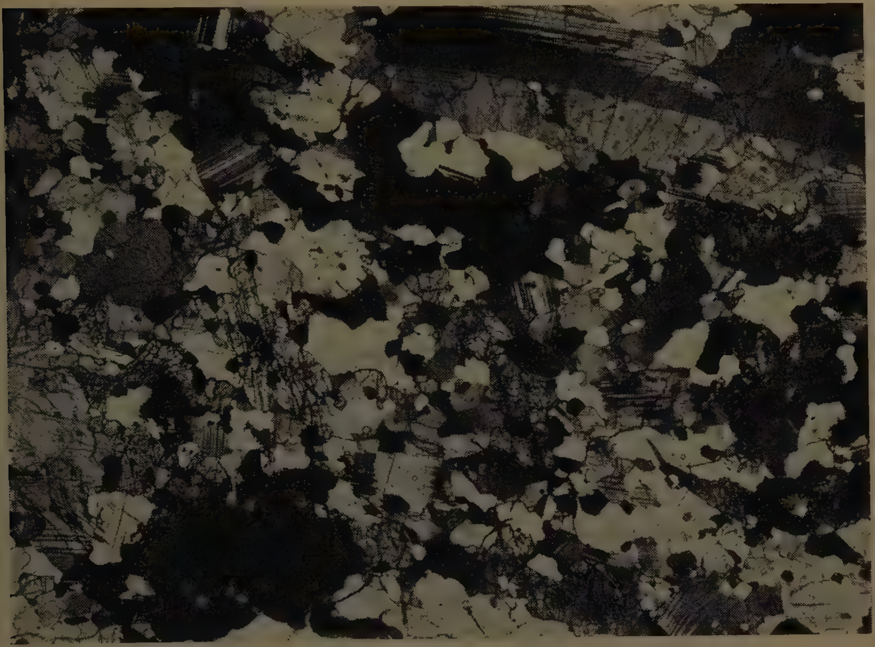


Fig. 5. Anatectic leuconorite of the Åseheia intrusive. Plagioclase of the first type (upper right corner) surrounded by the anatectic association in which hypersthene shows a poecilitique interstitielle texture (center). Nicols + ; 8x.

III. The Åmdal—Helleren—Rødland anorthosito-leuconoritic body.

The Åmdal—Helleren—Rødland body (Å—H—R) is the largest studied unit. It extends from north to south, between the Helleland region and Rekefjord, over approximately 20 km, and from west to east, between Åmdal and Barstad, over about 12 km.

On the west, it intrudes into the two units described above and reaches the neighbourhood of Åmdal (protuberancy of Åmdal). On the northwest, it cuts across the norito-granitic band and the Egersund—Ogna body; on the northeast, it forms protuberancies intruding the migmatitic band of the Lakssvelefjell and the leuconoritic and noritic horizons of the inferior part of the Bjerkreim—Sokndal body (protuberancies of Støla and Rødland); on the east, it is cut by the monzonorite of Eia which represents the last magmatic phase of the same body. Finally, on the south, it slopes down under the sea.

The contacts and the xenoliths indicate that the Å—H—R body is a magmatic unit with a well defined transgressive character. The central part is a coarse-grained anorthosite with sometimes very large euhedral crystals, containing in places patches of leuconorite; the border part is a medium-to coarse-grained leuconorite, which is also found in the protuberancies of Åmdal, Støla, and Rødland.

The relations between the leuconorite and the anorthosite are analogous to those found in the northern anatectic part of the Håland body: indeed, the association anorthosite-leuconorite is in both places characterized by the reciprocal relations between these two rocks. The leuconoritic border of the body contains coarse-grained anorthositic inclusions. In its central part the relative proportions of the two rocks become more or less equal and finally, in the center, the anorthosite forms a rather huge and homogeneous unit locally containing patches of leuconorite.

Petrographical study.

The component minerals and the texture of the leuconorite are identical to those of the anatectic rocks of the Håland body and of the Åseheia and Augendal intrusives: the plagioclase of the first type, sometimes rather coarse, is older than the associations formed by plagioclase of the second type, hypersthene of a poecilitique interstitielle texture, diopside and black ore. The coarse-grained anorthosite shows plagioclases of the first type, i.e. similar to those of the residual para-anatectic rocks of the Håland body.

Conclusion.

The formation of the anorthosito-leuconoritic body of Å—H—R also would seem to be connected with the leuconoritic anatectic process. The volume of rock here affected is very great, and the residual masses are very coarse. Indeed, in the central area of the body, and in the coarse-grained anorthositic inclusions contained in the leuconorite, plagioclases from 20 cm to 1 m long are common. They are of identical composition and never show any zoning. The physical conditions during the anatexis must have been very stable for a very long period of time.

The pristine substrate on which the anatectic processes have worked can only be surmised; it seems possible to relegate it to rocks of the same type as those of the masses of the Håland body, i.e. gneisses similar to the gneisses on the inward border of the Egersund—Ogna body, and also to leuconorites and norites forming the inferior part of the Bjerkreim—Sokndal body.

General Conclusion.

The geological and petrographical study of the anorthosito-leuconoritic complex of Håland—Helleren indicates that the deep zone of the earth's crust is the locale of a leuconoritic anatectic process of a regional extent: the basic palingenesis.

This process, in its initial phase, more or less modifies the structure and composition of the ancient rocks (Håland body); at greater intensities leuconoritic liquids are mobilized and large masses of para-anatectic anorthosite are formed (Åseheia and Augendal intrusives, Åmdal—Helleren—Rødland body).

The palingenetic phenomena started and developed later than the several stages of folding; they belong to the final stages of the regional plastic deformation.

The basic palingenesis throws new light on the problem of the origin of the large anorthositic masses. Indeed, the anatexis, in conjunction with the squeezing out of the dark components of a leuconoritic rock, accounts for the accumulation in large bodies of plagioclasic residues.

And so we come to a conclusion which corroborates the views expressed by P. Michot in his presidential address to the Société géologique de Belgique (P. Michot, 1956): «déjà effective dans la catazone profonde, l'intensité de ce phénoméne (l'anatexie leuconoritique) ne peut que croître dans les milieux plus profonds encore, pour produire, non seulement des magmas régénérés, mais encore et surtout pour engendrer de grandes masses anorthositiques de caractère paraanatectique. Ces derniers produits viendront ainsi s'ajouter aux anorthosites d'origine magmatique directe, pour constituer en profondeur, plus exactement dans les tréfonds de l'écorce terrestre ... de puissantes unités géologiques, douées d'une réelle individualité ...».

REFERENCES.

BARTH, T. F. W. 1936: The large pre-Cambrian intrusive bodies in the southern part of Norway. International Geological Congress; report of the XVI session, 1933; pp. 297—309; 1936.

— 1945: Geological map of the western Sörland. Norsk Geol. Tidsskr., Bd. 25, pp. 1—9, 1945.

GOLDSMITH, R. 1959: Granofels, a new metamorphic rock name. Jour. of Geol., Vol. 67, pp. 109—110, 1959.

KOLDERUP, C. F., 1896: Die Labradorfelse des westlichen Norwegens. I. Das Labradorfelsgebiet bei Egersund und Soggendal. Bergens Museum Aarbog, no V, 1896.

— 1914: Egersund Fjeldbygningen inden Rektangelkartet Egersund Omraade. Norges Geol. Undersökelse, no. 71, 1914.

MICHOT, J., 1958: Les plagioclases du massif anorthosito-noritique de Håland, Egersund (Norvège). Ann. Soc. Géol. de Belgique, T. LXXXI, pp. B 425—440, 1957—1958.

— 1959: Acception généralisée du terme cornéite. Ann. Soc. Géol. de Belgique; T. LXXXIII, pp. 87—89.

— 1960: La palingenèse basique. Académie royale de Belgique; Bull. Classe des Sciences, 5e série, T. XLVI, pp. 257—268.

— 1961 Le massif complexe anorthosito-leuconoritique de Håland—Helleren et la palingenèse basique. Mémoires de l'Académie royale de Belgique. Classe des Sciences, collection in—4°, 2e série, T. XV, fasc. 1, pp. 1—116, 1961.

MICHOT, P. 1939-a: La couronne d'anorthosite hypersthénifère feuilletée et rubanée du massif anorthositique d'Egersund (Norvège). Ann. Soc. Geol. de Belg., T. LXXII, pp. B 547—551, 1938—1939.

— 1939-b: La série rubanée norito-granitique du massif anorthositique d'Egersund (Norvège). Ann.Soc. Géol. de Belg., T. LXII, pp. B 532—546, 1938—1939.

— 1948: L'équilibre minéralogique dans les roches éruptives et le cadre géologique. Académie royale de Belgique; Bull. Classe des Sciences, 5e série, T. XXXIV, pp. 167—187.

— 1955-a: Anorthosites et anorthosites. Académie royale de Belgique; Bull. de la Classe des Sciences, 5e série, tome XLI, pp. 275—294, 1955.

— 1955-b: L'anatexie leuconoritique. Académie royale de Belgique; Bull. de la Classe des Sciences, 5e série, tome XLI, pp. 374—385, 1955.

— 1956: La géologie des zones profondes de l'écorce terrestre. Ann. Soc. Géol. de Belg., T. LXXX, pp. B 19—60, 1956—1957.

Manuscript received March 3, 1961.

Printed December 1961.

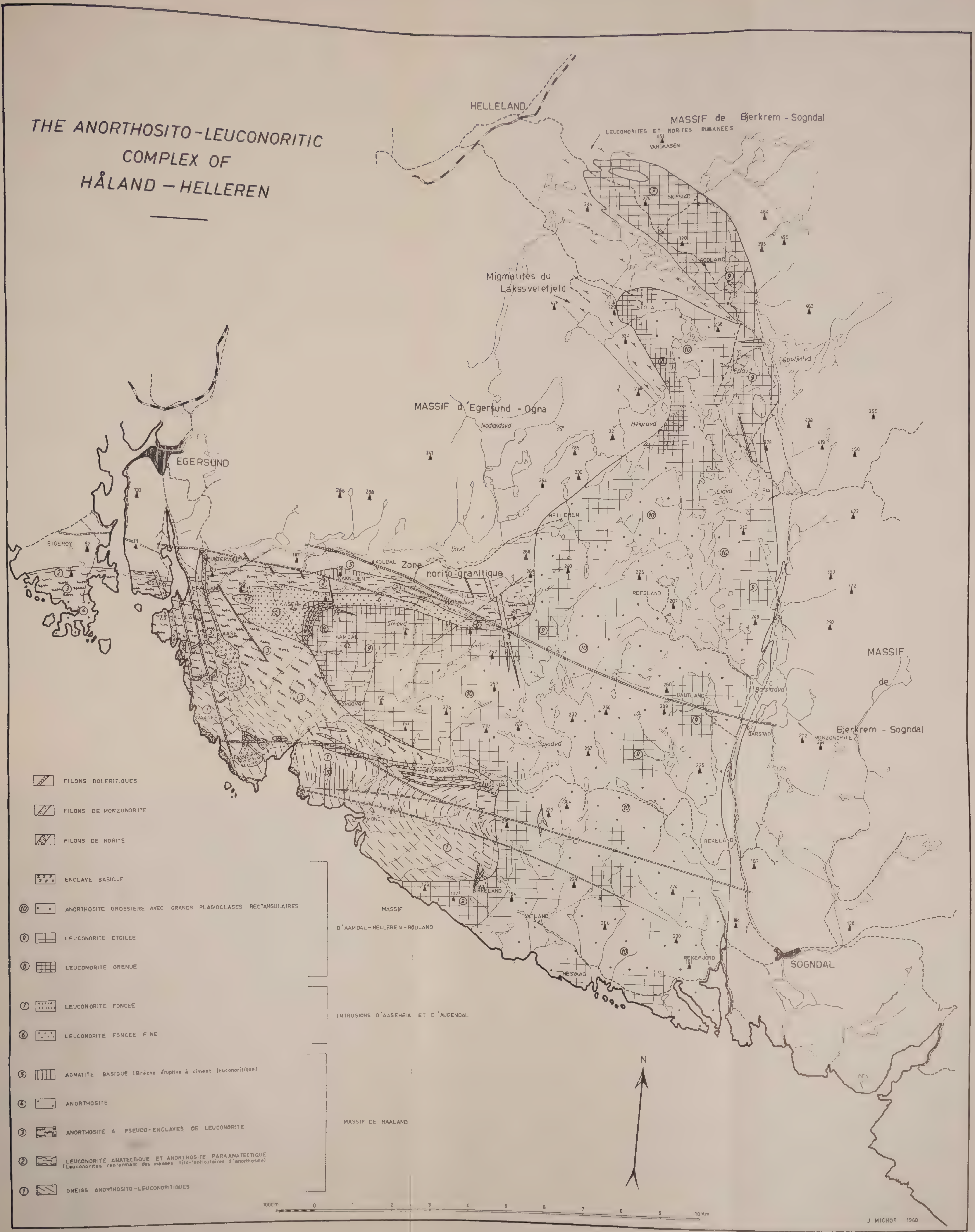
PLATE 1

Plate 1.

Geological map of the anorthosito-leuconoritic complex of Håland–Helleren

1–5: Håland body. — 1: anorthosito-leuconoritic gneisses (anorthositic granofels, leuconoritic gneisses and granofels, noritic gneiss); 2: anatectic leuconorite and para-anatectic anorthosite; 3: anorthosite with pseudo-inclusions of leuconorite; 4: anorthosite; 5: basic agmatite.
6, 7: Åseheia and Augendal intrusives. — 6: fine-grained leuconorite; 7: medium-grained leuconorite.
8–11: Åmdal—Helleren—Rødland body. — 8: coarse-grained leuconorite; 9: leuconorite; 10: coarse-grained anorthosite; 11: basic inclusions.
12: noritic dykes.
13: monzonoritic dykes.
14: doleritic dykes.

THE ANORTHOSITO-LEUCONORITIC COMPLEX OF HÅLAND – HELLEREN


HELLELAND
MASSIF de Bjerkrem - Sogndal
LEUCONORITES ET NORITES RUBANEES
VARDAASEN
SKIPSTAD
RODLAND
Migmatites du Lakssvelefjeld
STOLA
Grosfjellvd
Eptavd
MASSIF d´Egersund - Ogna
Nodlandsvd
Heigravd
EGERSUND
Eiavd
EIA
HELLEREN
EIGEROY
Liavd
AKOLDAL
KNUDEN
Zone norito-granitique
HESNES
HALLAND
AASE
AASEHEI
AAMDAL
Smievd
REFSLAND
NODLAND
SVAANES
STANNES
GAUTLAND
Barstadvd
BARSTAD
MASSIF de Bjerkrem - Sogndal
MONZONORITE
Spjodvd
Augendalvd
AUGENDAL
MONG
REKELAND
BIRKELAND
VATLAND
NESVAAG
REKEFJORD
SOGNDAL
FILONS DOLERITIQUES
FILONS DE MONZONORITE
FILONS DE NORITE
ENCLAVE BASIQUE
⑩ ANORTHOSITE GROSSIERE AVEC GRANDS PLAGIOCLASES RECTANGULAIRES
⑨ LEUCONORITE ETOILEE
⑧ LEUCONORITE GRENUE
MASSIF D´AAMDAL – HELLEREN – RÖDLAND
⑦ LEUCONORITE FONCEE
⑥ LEUCONORITE FONCEE FINE
INTRUSIONS D´AASEHEIA ET D´AUGENDAL
⑤ AGMATITE BASIQUE (Brèche éruptive à ciment leuconoritique)
④ ANORTHOSITE
③ ANORTHOSITE A PSEUDO-ENCLAVES DE LEUCONORITE
② LEUCONORITE ANATECTIQUE ET ANORTHOSITE PARAANATECTIQUE (Leuconorites renfermant des masses lito-lenticulaires d´anorthosite)
① GNEISS ANORTHOSITO-LEUCONORITIQUES
MASSIF DE HAALAND
N
1000 m
0
1
2
3
4
5
6
7
8
9
10 Km
J. MICHOT 1960

STRUKTURELLE IAKTTAGELSER VED NOEN NORSKE KISFOREKOMSTER

Structural observations on some Norwegian sulphide deposits

Av

HANS-PETER GEIS

Abstract. A number of deposits of sulphide ores in Norway are to be found in minor basins in the Caledonian geosyncline. The ore deposits are separated into cigar-shaped ore bodies which are arranged "en echelon". Their relationship to the basins is described and an explanation for the relationship is given.

Summary

The author gives a summary of his observations on Caledonian sulphide ore deposits in Norway. On the west coast the copper bearing pyrite mines of Visnes are to be found in the deepest parts of a trough filled with sedimentary and lava greenstones. At present this trough forms a syncline. Old Visnes mine (exhausted) consisted of about 6 cigar-shaped ore bodies, the greater axes of which were parallel to the dip of the surrounding rock. The ore bodies were arranged "en echelon" with the western neighbour always lying to the north. New Visnes mine (Rødklev) is crossed by a fault. The western part is a simple cigar-shaped body, the eastern part is very complicated because of many faults parallel to the strike. The author believes that the syncline continues in the Hardangerfjord where several plate-like ore bodies are to be found. In contrast to Visnes the longer axis of some ore bodies here does not always lie parallel to the lineation in the surrounding rocks. In western Norway pyrite deposits lying within lava greenstones are usually cigar-shaped, while those lying within sedimentary greenstones are more plate-like.

In central Norway there are the deposits of Hjerkinn and Folldal. Approaching the Hjerkinn pyrite deposit one observes the thickness of the surrounding rocks (greenstones) to increase. In the Folldal region there is a trough similar to the one on the west coast. This trough was discovered by geophysical investigations. Here the copper- and zinc-bearing pyrite ores appear not only in the center of the trough, but also on its flanks, Cu-contents increase towards the center. The ore bodies are plate-like and show an arrangement "en echelon" that corresponds to that of the zones of geophysical anomaly.

Bleikvassli mine in northern Norway is a lead and zinc bearing pyrite deposit with similar arrangement "en echelon" of the ore plates as those described before. The deposit is situated below the thickest part of a lens-shaped limestone layer.

From the descriptions of other authors one can see that similar structures are also to be found other places in Norway.

Finally the author gives the following explanation for the arrangement "en echelon". He supposes a submarine precipitation of hydrothermal solutions in the deepest of the elongated basins. Together with this precipitation deposition of sedimentary and volcanic material takes place and thus the basins are filled little by little. But there will be deposited relatively more material in the vicinity of the place of origin of the material. This way the axis of the basin wanders and with it the precipitation of the sulphides.

Ved undersøkelser på kismalmforekomster i Norge har forfatteren søkt etter lovmessigheter som kunne lette malmletingen og gruvedriften. I det følgende skal det berettes om noen trekk som går igjen ved en rekke kisforekomster. Det er i stor utstrekning egne iakttagelser som meddeles, men det refereres også til tidligere utførte arbeider av andre forfattere. Jeg takker prof. Chr. Oftedahl, geolog F. J. Skjerlie og geolog M. Gustavson for kritisk gjennomsyn og kommentarer til manuskriptet.

Tidligere undersøkelser

En rekke forfattere har tidligere fått frem en del lovmessigheter. J. H. L. Vogt (1910) fant ut at kismalmen var linjalformet, ofte var tilknyttet gabbroaktige bergarter (sml. Vogt 1905. Falkenberg

1914) og at den lå konformt med lagene. FOSLIE (1926) og CARSTENS (1932) fant senere ut at kismalmene lot seg inndele på grunnlag av kisens tilknytning til forskjellige bergarter (Leksdal-, Røros-, Rødhammer-, Fløttumtypus). Også tilknytningen av kisforekomstene til de kaledonske fjellkjedeområdene på Vestlandet og i Trondheimsfeltet har jo vært kjent lenge og er nevnt av en hel rekke forfattere. I det siste har OFTEDAHL (1958) påpekt den betydning som sure tuffer har i forbindelse med kis- og jernmalmforekomster.

Når det gjelder andre strukturelle trekk så kan nevnes at REUSCH allerede i sin avhandling «Bømmeløen og Karmøen med omgivelser» fra 1888 konstaterer at malmstokkene i gamle Visnes gruve og Valaheien på Varaldsøy har sin lengdeakse parallell med en utpreget lineasjon i fjellet omkring. Siden ble det kjent at de fleste norske malmforekomster har en slik stokkform (VOGT 1910, CARSTENS 1935, VOGT 1948) og FOSLIE (1926) gir i «Norges svovelkisforekomster» en hel rekke eksempler på dette og nevner også «stjert-om-stjert»oppbygning i denne forbindelse. Dessuten ble det funnet ut at det i forbindelse med drivverdige kisforekomster gjerne opptrer langstrakte lag med impregnasjon av svovelkis, magnetkis, leilighetsvis også kobberkis, og man skilnet mellom «vasskis» og «gangkis». Dette ble bl.a. pekt på av VOGT (1910) og C. BUGGE (1949). CARSTENS (1935) henviser til at kislegemenes form alltid er en funksjon av tektonikken i det respektive område, d.v.s. at deres lengdeakser går parallelt med foldningsaksene i fjellet omkring. TH. VOGT (1945, 1948) som i en årrekke beskjeftiget seg med disse problemene, sier at dette ikke er tilfelle. Ifølge hans observasjoner går malmstokkenes lengdeakse parallelt med metamorfe flytestrukturer: biotitt-, magnetkis- og hornblendekrystaller er langstrakt i flyteretningen av bergartene under metamorfosen. Og denne retning er ifølge ham uavhengig av foldningsaksene, skjønt begge to *kan* ha samme retning. Han nevner Nord-Rana, Røros, Sulitjelma og Folldal som eksempler.

I den senere tid har VOKES (1957) funnet en viss «structural control» av malmen i Birtavarre: Mineraliseringen er her knyttet til et sted i Kåfjordsynklinalen hvor lineærstrukturene viser en markert sving fra NW—SE- til E—W-retning.

Kisforekomstene på Vestlandet

Jeg vil ta mitt utgangspunkt i Karmøy. Den geologiske bygning på Karmøy og langs kysten ved Haugesund er tidligere undersøkt av Reusch (1888), dessuten er det nevnt noe hos Goldschmidt (1921) og Isachsen (1940).

Forfatteren har på Karmøy og det tilstøtende fastland utført geologisk kartleggingsarbeide, i enkelte områder meget detaljert på flyfoto 1:5000 (en publikasjon som omhandler dette, er under forberedelse). Det viste seg herved at det på Karmøy kunne skilnes mellom to typer grønne bergarter:

En tynnspaltende grønn skifer med delvis utpreget lagdeling som av den grunn oppfattes som en sedimentær grønnsten, og en mer kompakt grønn bergart, hvori det på en rekke steder opptrer putestrukturer og andre tegn på flytende lava (lava-grønnsten). Herunder følger en bred sone med en blanding av fin- og grovkrystallinske bergarter som jeg med støtte i detaljobservasjoner oppfatter som en sterkt metamorfosert lavagrønnsten.

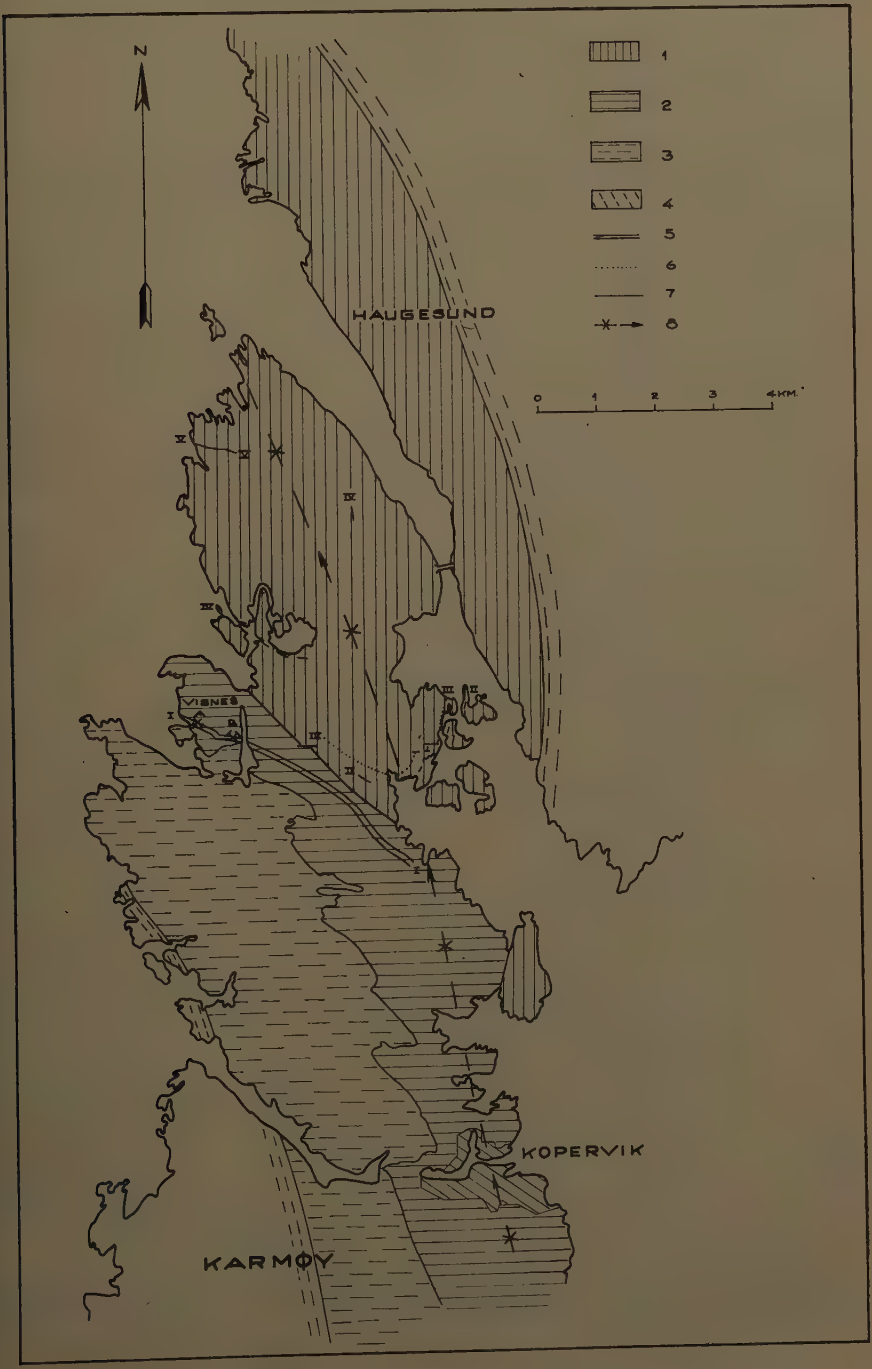
Reusch (1888) kalte den sedimentære grønnsten for «grønne, krystallinske skifere». Derimot slo han sammen lavagrønnstenen og

---

Fig. 1. Geologisk kart over den nordlige del av Karmøy samt omkring Haugesund. Etter egne resultater og etter Reusch.

Geological map of the northern part of the island of Karmøy and the district around Haugesund.

1 sedimentær grønnsten	*sedimentary greenstone*
2 lava-grønnsten	*lava greenstone*
3 "saussuritgabbro"	*"saussurite gabbro"*
4 gneis	*gneiss*
5 Visnes-skiferhorisont	*layer of greenschist*
6 grønnskifer med kvartslag	*greenschist with quartz layers*
7 mineraliserte soner	*mineralized zones*
8 synklinalakse med stupning	*axis of syncline with plunge*
I—I Visnes-malmhorisont	*ore zone of Visnes*
A Gamle Visnes gruve	*old Visnes mine*
B Rødklev gruve	*new Visnes mine*
II—II malmsone Våge—Gloppe	*Våge—Gloppe ore zone*
III—III grønnskifer med kvartslag	*greenschist with quartz layers*
IV—IV malmsone Vikingstad—Hauge	*Vikingstad—Hauge ore zone*
V—V Dale magnetkisdrag	*Dale pyrrhotite layer*


N
1
2
3
4
5
6
7
8
0
1
2
3
4 KM.
HAUGESUND
VISNES
KOPERVIK
KARMØY

de grovkystallinske bergartene og kalte dem for «dioritisk bergart». De grovkrystallinske partiene ble av GOLDSCHMIDT (1921) kalt for «Saussuritgabbro».

Lavagrønnstenen blir på holmene utenfor østkysten av Karmøy mer og mer tynnplatig-skifrig og på fastlandet omkring Haugesund finner man bare sedimentær grønnsten. Av den brede sone med grovkrystallinske bergarter er her bare en forholdsvis smal stripe igjen.

Da lagdelingen og de geologiske grenser lett lot seg fastlegge, var det ikke vanskelig å bestemme den tektoniske oppbygning. Bergartene er foldet på en slik måte at de danner en synklinal langs Karmøys østkyst (sml. KOLDERUP 1931). Dens akse faller med en liten vinkel mot NNW. På dette grunnlag var det også mulig å bestemme mektigheten av de 3 forannevnte grønne bergarter. Ved Haugesund er denne ca. 2 500 m, ved Visnes-forekomsten sammenlagt ca. 6 000 m. Den av KOLDERUP (1931) nevnte glimmerskifer i Torvastad har ingen innflytelse på den synklinalformete oppbygning på Karmøy.

Under de nevnte grønne bergarter opptrer det på fastlandet og på Sør-Karmøy høymetamorfe bergarter, overveiende injeksjonsgneis, men også flere andre gneistyper, samt granitt. Disse bergarter er tydeligvis foldet samtidig med de grønne bergarter. Deres stratigrafiske stilling og deres metamorfosegrad tyder derimot på en høyere alder, som jeg antar er omtrent den samme som den HERNES (1956) antok for tilsvarende bergarter i Molde—Kristiansund-området, d.v.s. prekambrium.

Malmen på Visnes ligger i et sedimentært lag (Visnesskifer, delvis kartlagt av REUSCH 1888) innen lava-grønnstenen. Mineraliseringen begynner 750 m sørøst for Rødklevsjakten og er, resp. var sterkest i Rødklev-gruve og gamle Visnes-gruve. Vestenfor ligger havet, så man vet lite om de geologiske forhold lenger mot NW. En blyantformet kisanrikning med 2 m mektighet ligger rett N for gruvene i Visnes (Haugesundsynken). 1,5 km NNE for Visnes er en lagformet forekomst av magnetitt. Den største anrikning ligger rett N for gruvene i Visnes. Enda lenger nord (4,2 km N for Rødklev-sjakten) finnes enda en lagformet mineralisering, her av magnetkis. *Vi ser altså her en viss tilknytning av de største malmmektighetene til områder med store grønnstensmektigheter.*

La oss nå se på de spesielle forhold i gruvene på Visnes. I gamle Visnes gruve opptrådte det 6 lansettformete stokker som etter strøket lå stjert-om-stjert på en slik måte at den neste mot NW var forskjøvet mot hengen (se fig. 2). Mektigheten gikk opp til 25 m. Beskrivelser er gitt av REUSCH (1884, 1888) J. H. L. VOGT (1894, 1910), FOSLIE (1926), DITTMARSCH-FLOCON (1875), HELLAND (1871), KNUDSEN (1885). Forholdet mellom maksimalmektighetene og stokkenes bredde var lite. Man hadde altså ikke den typiske «linjalform», men mere «brødform». Noe liknende gjelder for Rødklev-gruven. Her har man malm på to sider av den N—S-gående Visnesvann-forkastning, «Vestkisen» og «Østkisen». Vestkisen — FOSLIE (1926) har beskrevet den — kan nærmest kalles for en linjal, men en meget uregelmessig en, med en rekke fortykkelser og forgreninger. Østkisens oppbygning er langt mer komplisert. En rekke parallelt med strøket gående forkastninger har skjøvet kisen sammen nærmest som et trekkspill og det har hittil ikke vært mulig å rekonstruere malmlegemets opprinnelige form (se fig. 3). Det kan hende at også dette tidligere hadde en brødliknende form. Dessuten kan visse observasjoner tyde på at malmen har utkilinger mot hengen og mot dypet. Av de gamle publikasjoner fremgår det at man i gamle Visnes gruve hadde utkilinger i motsatt retning.

Forholdene på Stord er undersøkt av REUSCH (1888), KVALE (1937) og SKORDAL (1950). Ved befaringer fikk forfatteren inntrykk av at det forelå en tilsvarende bergartsfølge som på Karmøy, men strøkretningen er her SW—NE. Litlabø gruve ligger i de bergarter som tilsvarer den sedimentære grønnstensavdeling på Karmøy. Malmforekomsten er jo et av de peneste eksempler på linjal som finnes i Norge (se SKORDAL 1950 og JONSSON 1952). Ved befaringene fikk forfatteren også inntrykk av en liknende synklinal-oppbygging som på Karmøy. Den antatte synklinalakse går fra sørspissen av Bømlo, Stord, Tysnes og på Ølve over til fastlandet. Men jeg vil gjerne nevne at prof. Kvale under en diskusjon ikke fant å kunne være enig i dette. Lengre mot NE ligger — også i sedimentær grønnsten, men på den andre siden av den antatte synklinal — forekomstene på Ølve og Varaldsøy (sml. REUSCH 1888 og FOSLIE 1926, 1955). Det ser ut til at forholdene her er noenlunde de samme som ved Haugesund: bare sedimentær grønnsten på østsiden, større mektigheter med lavagrønnsten på vestsiden.

I de gamle Atramadalsgruvene på Ølve opptrer malmen nesten som en fløts med mektigheter opp til 1—2 m. Bare Gravdal gruve er f.t. i drift og her opptrer det igjen en typisk linjalform, nærmere be-

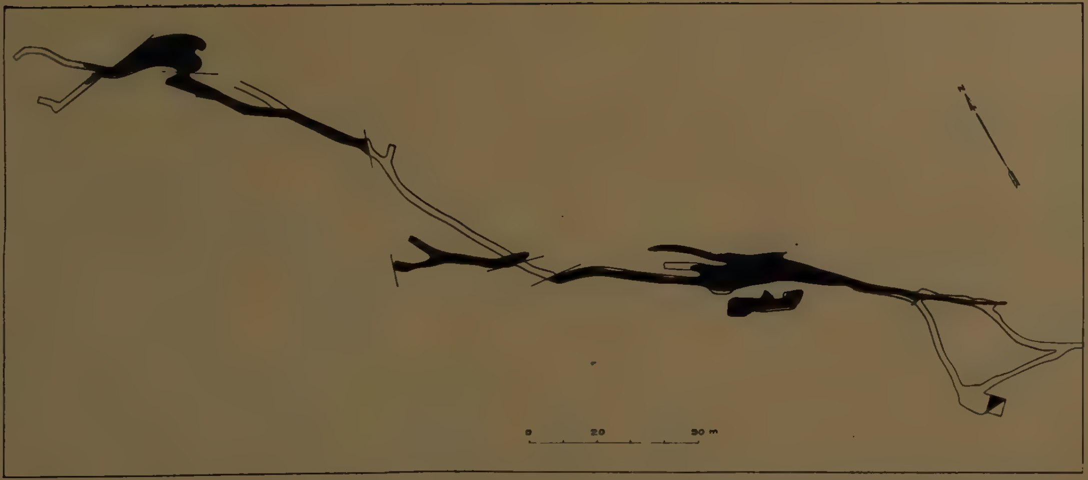


Fig. 2. Horisontalsnitt gjennom etasje 160 m i gamle Visnes gruve. Sort: malm. Etter gammelt gruvekart.

Horizontal section on the 160 m level, old Visnes mine. Black: ore.

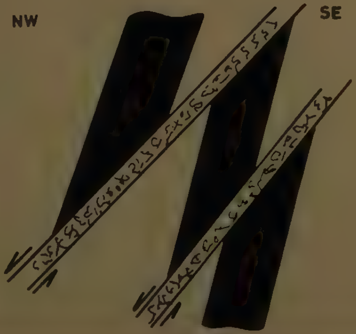


Fig. 3. Profilsnitt gjennom østkisen i Rødklev gruve. Prinsipp-skisse. Sort: malm.

Vertical section through the eastern part of the new Visnes mine. Generalized. Black: ore.

stemt to i det vesentlige parallelle linjaler som stuper med 20° mot SW. Interessant er en observasjon som driftsbestyrer Meisingset har gjort: der hvor malmen kiler ut, er den gjerne foldet og krøllet. Såvidt jeg har kunnet se under befaringen, er lineasjonen horisontal.

Valaheien kisgruve på Varaldsøy som etter krigen ble overtatt og åpnet av Stordø kisgruber er i de øvre partiene en ganske regelmessig linjal. Men i den nyoppfarete stoll har det vist seg at linjalens lengdeakse her har en annen retning. Gruven er forresten et godt eksempel på «linjal» parallelt med den tektoniske strekningsretning som allerede Reusch (1888) gjorde oppmerksom på. Foslie (1926) nevner dessuten stjert-om-stjert-oppbygning.

Sammendrag (Vestlandet)

I Visnes er malmforekomstene knyttet til det sted hvor grønnstensmektigheten er størst. I Hardangerfjorden er det p.g.a. de flate akseretninger ikke så lett å si om en slik strukturell tilknytning også gjelder for forekomstene her. Derimot kan man si: Forekomstene i sedimentær grønnsten har linjal- til skiveform og er forholdsvis regelmessig utformet. Forekomstene i lavagrønnsten har en mere brødliknende form og er mindre regelmessige. Lineasjonen i sidebergarten behøver ikke gå parallelt med malmstokkenes lengdeakse og akseretningen kan forandre seg mot dypet.

Kisforekomster på Hjerkinn og i Folldal

Geologien i Hjerkinn-området er ikke særlig godt kjent, men det er utført en del undersøkelser i områder omkrig. DuDresnay (1950) og Holmsen (1955) har arbeidet i nord, Vogt (1889) og Marlow (1935) i øst. Prof. Strand har — såvidt vites — utført en del kartlegging på Dovre, men det er ikke publisert noe.

Ved forfatterens undersøkelser viste det seg at det i store trekk opptrer denne lagfølge:

Øverst: fyllitter (tilsv. antakelig Hovin-gruppen i Trondheimsfeltet)
grønnstener med sedimentære lag (antakelig tilsv. Bymarkgruppen).
Nederst: glimmerskifer (antakelig tilsv. Rørosgruppen).

Denne lagfølge finner man ved Hjerkinn med strøkretning nesten W—E og fall meget steilt mot S. Mot Snøhetta kommer man ned i eldre bergarter. I Folldalsområdet har den samme lagfølge strøkretning SW—NE og fall steilt mot NW. Her kommer man i eldre bergarter når man nærmer seg Grimsdalen. Det foreligger altså en bueformet synklinal med konkavsiden mot NW. Den stemmer forøvrig meget godt overens med VOGTS synklinal II (VOGT 1954).

Hjerkinn- eller Tverfjellforekomsten ligger 2 km NW for Hjerkinn stasjon i den steiltstående nordsjenkel av den ovennevnte synklinal og innen grønnstenene med strøkretning E—W. Disse grønnstener inneholder også rent sedimentære innleiringer. Mens malmforekomsten ligger i et område hvor man nesten utelukkende finner tynnskifrige vulkanske bergarter, dominerer i samme geologiske horisont såvel øst som vest for malmen det sedimentære innslag med garbenskifer, kvartsitt, kvarts-serisitt-skifer og fyllitt. Først øst for Hjerkinn Fjellstue — omkring Brendhøin — overveier igjen vulkanske bergarter. Ved elektromagnetiske målinger oppdaget Geofysisk Malmleting en 10 km lang anomalisone (Rapport 103/1953). Den viser en slik stjert-om-stjertoppbygning at dens avstand fra Bymarkgruppens henggrense blir større når man nærmer seg Hjerkinnmalmen.

Selve malmforekomsten er boret opp inntil 150 m dyp med ialt 29 diamantborhull. Herved har det vist seg at den består av 3 typiske malmlinjaler hvis lengdeakse stuper med 30—45° mot øst (se fig. 4). Lineasjonene i fjellet omkring derimot går i horisontal retning som fig. 5 viser. Om andre detaljstrukturer kan man enda ikke si noe.

Folldal-malmforekomstene ligger i *sør*sjenkelen av den samme synklinal, den har her strøkretning WSW-ENE. Malmen ligger også her i grønnstener. Dette er de av MARLOW (1935) nevnte hornblenderike skifere. Her er det av Geofysisk Malmleting utført geofysiske målinger på en strøklengde av ca. 20 km (kart fra 1957), se fig. 6. Herved ble det funnet en nordlig og en sørlig anomalisone. Helt i vest — omkring Grimsdalsgruven — ligger anomaliene tett i tett, østover forsvinner noen av dem til det bare er to hovedanomalidrag igjen som fjerner seg mer og mer fra hverandre jo lengre øst man kommer, samtidig som de splitter seg opp. Det nordlige anomalidrag er ikke blitt fulgt lenger enn til rett nord for bebyggelsen i Folldal. Den sørlige er — i alle fall delvis — blitt fulgt til 3 km NE for nordre Gjetryggen gruve (Rapport fra 1941).

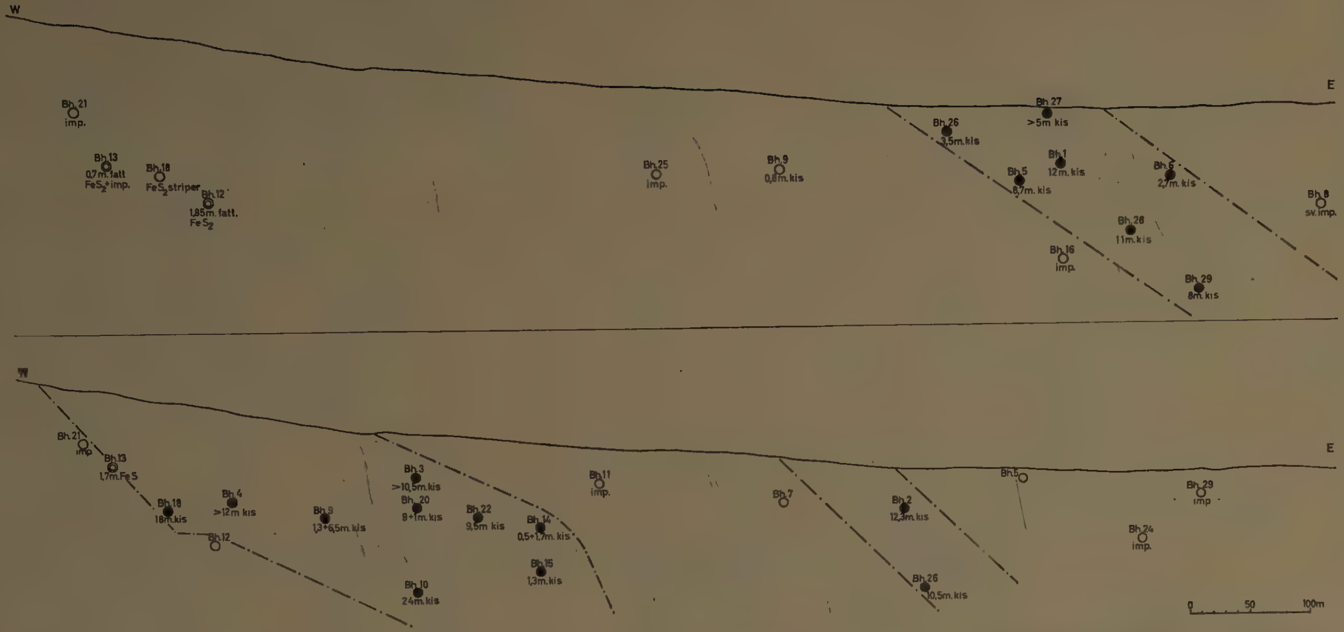


Fig. 4. Vertikale lengdesnitt gjennom malmene på Hjerkinn. Tallene oppgir mektighetene påtruffet i borhullene. Avstand mellom de to profillinjene ca. 50 m.

Vertical section through the Hjerkinn ore bodies. Bh. = drillhole with thickness of the ore. Distance between profiles about 50 m.

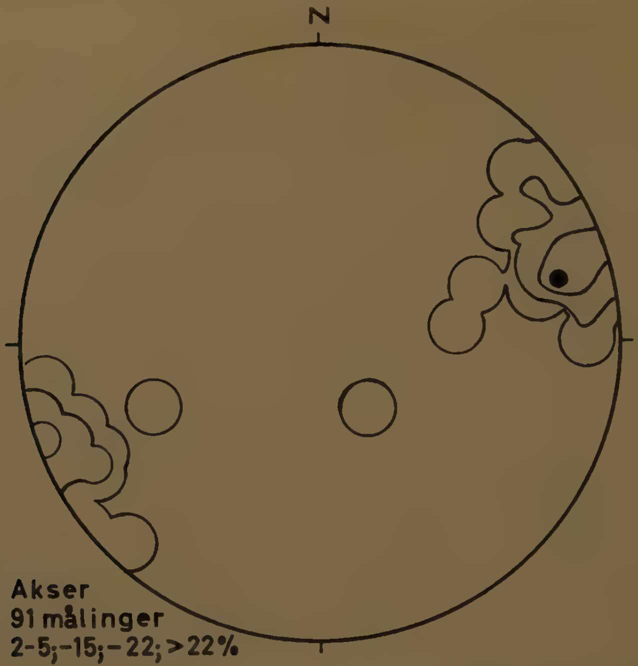


Fig. 5. Statistisk diagram av lineasjonene omkring Hjerkinnmalmen.
Statistical diagram of linear structures around the Hjerkinn ore bodies.

Det mest iøynefallende ved anomaliene er buene som de danner og det faktum at de splitter seg opp. Det må vel tydes slik at bergartspakkene mellom dem blir tykkere, d.v.s. som en primær mektighetsøkning østover. Stjert-om-stjert-oppbygning finner man også noen steder, meget utpreget er den omkring nordre Gjetryggen gruve (GM-rapport fra 1941). Her finner man fortsettelsen av anomalien østover forskjøvet mot nord. I det målte område som ligger vestenfor gamle Folldal hovedgruve, kan man se antydning til stjert-om-stjert-oppbygning noen steder med motsatt anordning, d.v.s. fortsettelsen mot vest er forskjøvet mot nord, men dette er ikke så utpreget.

La oss nå se litt nærmere på malmforekomstene: Vi går fra vest mot øst. Grimsdalsgruvens malm er antakelig en meget regelmessig skive som er kjent over 1500 m lengde og mektigheten er på 1—2 m. Dragningen i felt antas å være ca. 30° mot vest. Cu-gehaltene ligger under 1% og av jernsulfidene er 40% magnetkis (Hjelseth &

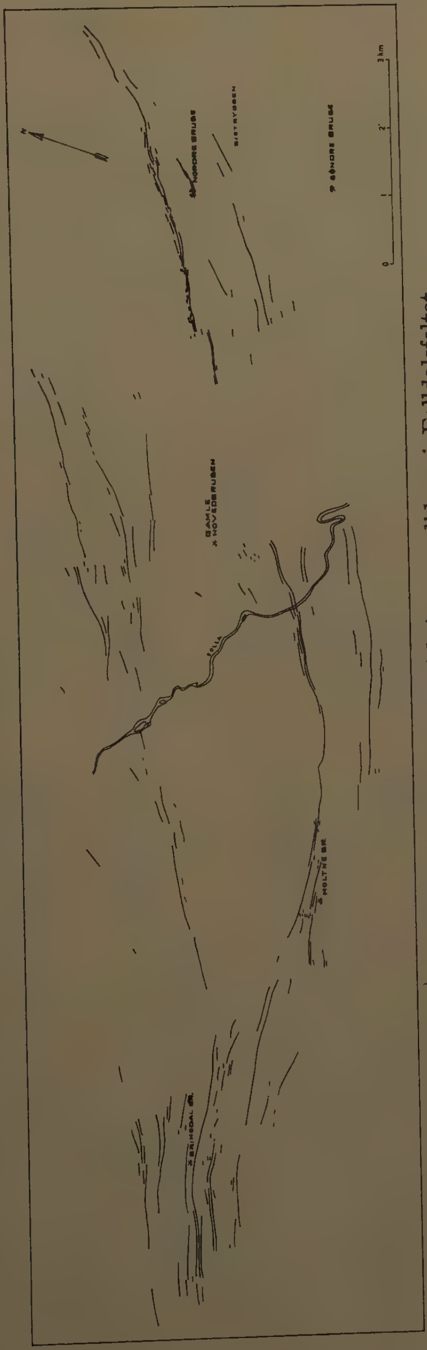


Fig. 6. Geofysiske (elektromagnetiske) anomalidrag i Folldalsfeltet.
Zones of geophysical (electromagnetic) anomaly in the Folldal district.

EINARSEN 1950). Forekomsten ligger i et område med 10 parallelle anomalisoner.

Grev Moltke- eller Nygruven (FOSLIE 1926, MARLOW 1935, HJELSETH & EINARSEN 1950) var en typisk linjalforekomst med aksefall 12—15° mot vest, en bredde av 50 m og en mektighet mellom 1 og 8 m. Ved et tverrsprang var den vestlige del forskjøvet ca. 130 m nedover. Råmalmen inneholdt omkring 1% Cu og mindre magnetkis enn Grimsdalsmalmen. Forekomsten lå litt sør for det sørlige anomalidraget og skiller seg ut ved at det her finnes 4 korte anomalisoner.

Akkurat i det område hvor gamle Folldal hovedgruve lå, mangler det geofysiske målinger. Det ser imidlertid ut til at den lå ca. 500 m N for det sørlige anomalidrag. Gruven er beskrevet hos FOSLIE (1926), J.H.L. VOGT (1889) og MARLOW (1935). Ifølge disse var den malmførende sone kjent over en strøklengde av ca. 900 m. Det var tre linseaktige malmlegemer, de enkelte lengdeakser hadde stupning mot vest. Fra SW til NE gikk linsene dypere slik at akseretningen av linsenes anordning forløp omtrent loddrett på hver enkelt linses lengdeakse, den sentrale linse var den avgjort største. Av FOSLIES gruvekart får man inntrykket av en viss stjert-om-stjert-oppbygning: Den østlige linse var i forhold til sentrallinsen forskjøvet litt mot hengen. Malmen var en meget ren svovelkis med gjennomsnittlig ca. 2% Cu.

Nordre Gjetryggen gruve, Folldals hovedmalmkilde f.t., ligger ca. 500 m sør for det sørlige anomalidrag. Forekomsten kan kalles en «linseskare» (sml. VOGT 1948) og viser således en liknende oppbygning som den gamle gruvens malm. Også her varierer linsenes størrelse. De enkelte linsers lengdeakser strekker seg i nordvestlig retning, mens linseskarens lengdeakse faller mot NE (HJELSETH & EINARSEN 1950). Men linsene er ikke enkle skiver, de er forholdsvis uregelmessige dannelser med fortykkelser, forgreninger og innleiringer av sidebergart osv. Under gruvedriften oppdaget man at forgreningene skjer etter en viss lovmessighet: fra hengsiden av malmen går malmkiler i vestlig retning, fra liggsiden av malmen går malmkiler i østlig retning (se fig. 7). Dessuten har man noen steder funnet at malmen deler seg også mot dypet. Liggmalmen kiler da ut etter en stund, mens hengmalmen fortsetter. Dette bevirker at malmen mot øst og mot dypet kommer til å ligge i yngre lag. Det skal bemerkes at dette tilsvarer tendensen i stjert-om-stjert-oppbygningen av de geofysiske

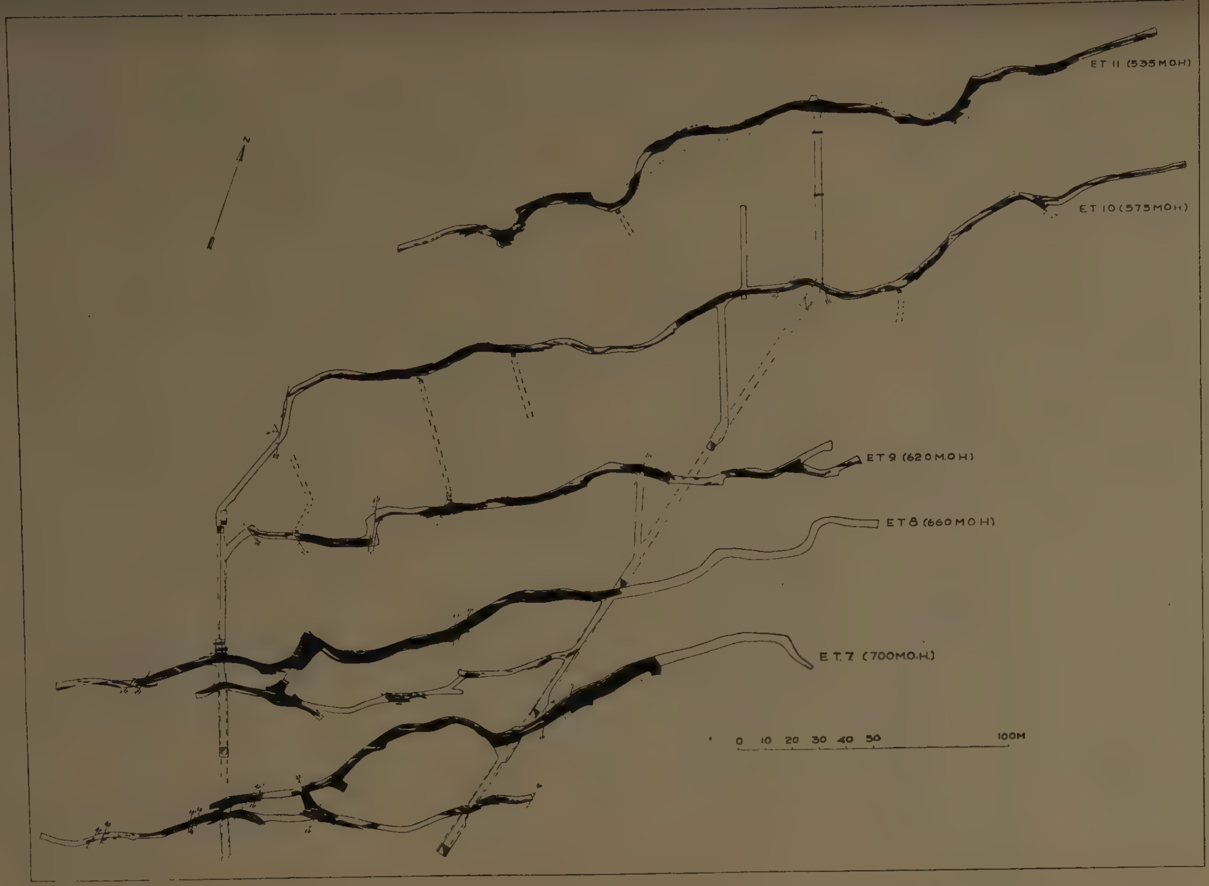


Fig. 7. Malmen på etasjene 7—11 i nordre Gjetryggen gruve, Folldal. Sort: malm.
Horizontal section on level 7—11, nordre Gjetryggen mine, Folldal. Black: ore.

anomalier. Cu-gehalten av malmen er ca. 1%, magnetkisgehalten minimal.

Forekomsten søndre Gjetryggen gruve ligger 2 km sør for nordre Gjetryggen gruve og dermed også temmelig nøyaktig rett under den. Også den består av en rekke enkeltlinser, men bortsett fra den sentrale linse er de små og man har foreløpig ikke klart å få overblikk over deres anordning og over fortsettelsen mot dypet. Man finner forgreninger av malmen: Fra hengsiden stikker korte kiler ut mot øst, fra liggsiden mot vest. Det er altså det omvendte av forholdene i nordre Gjetryggen gruve. Malmen holder litt under 1% Cu. Eldre beskrivelser foreligger av FOSLIE (1926), MARLOW (1935), HJELSETH & EINARSEN (1957).

Sammendrag (Folldal–Hjerkinn)

Jeg skal nå gjøre et forsøk på å vise de store trekk ved malmforekomstene i Hjerkinn—Folldal-området. Det er forsåvidt en viss likhet med Karmøy—Hardanger-området som det også her er malmforekomster i begge sjenkler av en synklinal i grønnsten. Grønnstenen er her overveiende tynnskifrig og forekomstene som ventet skive-til linjalaktige. Folldal-forekomstene ligger i det vesentlige i nærheten av det sørlige geofysiske anomalidrag som finnes der. Bortsett fra søndre gruve er deres avstand fra hovedanomalidraget opp til 500 m. Grimsdalsgruven og gamle Folldal hovedgruve ligger nord for det, Grev Moltke, nordre og søndre Gjetryggen ligger sør for det. Langs anomalidraget er den innbyrdes avstand mellom gruvene 4—5 km. Kobbergehaltene øker i retning av gamle Folldal hovedgruve såvel fra øst som fra vest (Grimsdalsgruven under 1%, Grev Moltke omkring 1%, Hovedgruven omkring 2%, nordre Gjetryggen omkring 1%), i samme retning avtar magnetkisgehalten. Vi merker oss at kobbergehalten er størst hvor også avstanden mellom de to anomalidragene er størst. Grimsdalsgruven og Grev Moltke ligger dessuten på steder hvor de langstrakte anomalidragene viser uregelmessigheter. Hvor disse uregelmessigheter mangler i forbindelse med forekomster, er de anomaliene som tilsvarer malm forholdsvis korte og ligger enten på ligg- eller hengsiden av hoveddraget. Til slutt skal det nevnes at malmens forgrening og stjert-om-stjert-oppbygning og anomalienes stjert-om-stjert-anordning i nordre Gjetryggen gruve faller sammen.

Bleikvassli

Bleikvassli Gruber driver med sin råmalm på 5,5% Zn, 3,0% Pb og 22% S i grunnen ikke på en typisk kisforekomst. Men malmlegemets oppbygning likner såpass på de før nevnte kisforekomster at den her også skal nevnes (se fig. 8). Malmen er temmelig komplekst sammensatt, den består av svovelkis, magnetkis, blyglans, sinkblende og litt kobberkis. I store trekk er det en skive.

På den øverste etasje er malmen i det vesentlige sammenhengende med utkilinger mot hengen og liggen etter de samme prinsipper som i nordre Gjetryggen gruve i Folldal. Men det er en viss forskjell: I den nordlige halvpart av den øverste etasjen går utkilingene mot hengen i sørlig retning, utkilingene mot liggen går i nordlig retning. I den sørlige halvpart er det omvendt. Oppbyggingen av det horisontale malmtverrsnitt er altså noenlunde symmetrisk. Men symmetriaksen faller med en meget flat vinkel i sørvestlig retning, slik at det på etasjene under — de ligger rett under hverandre — bare opptrer utkilinger mot hengen i sørlig retning og utkilinger mot liggen i nordlig retning. Om utviklingen etter vertikalplanet kunne det ikke gjøres sikre observasjoner.

De geologiske forhold har vi noe kjennskap til på grunnlag av REKSTADS geologiske generalkart, blad Hatfjelldal (REKSTAD 1924). Av dette fremgår det at forekomsten ligger i et område med kaledonske glimmerskifre og kalkstener, på østsiden av et kalkstensdrag som faller mot vest. Kalkstensdragets mektighet varierer sterkt og forekomsten ligger på liggsiden temmelig nøyaktig ved den største kalkstensmektighet.

Andre forekomster i Norge

Jeg skal nå foreta en sammenlikning med undersøkelser som andre geologer har utført og med de praktiske driftserfaringene fra andre steder.

Et felt som er meget godt kjent er Grongfeltet hvor FOSLIE (1957, 1958 a-c) har utført geologisk kartlegging og FOSLIE (1926) og OFTEDAHL (1958) malmgeologiske undersøkelser. Av deres publikasjoner får man følgende bilde: Der opptrer en mengde forekomster, de fleste er små og uten økonomisk betydning, men 3 av dem er av større dimensjoner: Skorovass, Joma og Gjersvik.

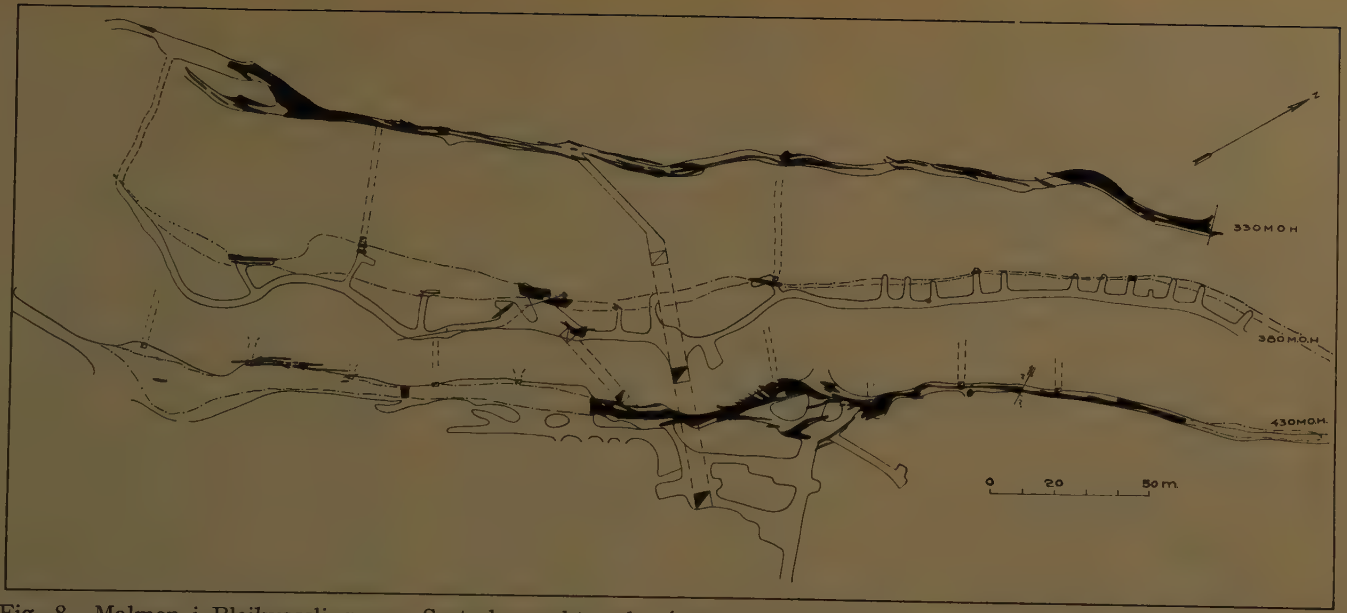


Fig. 8. Malmen i Bleikvassli gruve. Sort: kompakt malm (eventuelt drivverdig impregnasjon er ikke tatt med). Strek-prikk-linjet: begrensning av de utdrevne partier. Kartet viser det som var synlig november 1958. I mellomtiden ble det funnet nye malmpartier på 330 m-etasje, de største partiene på kartet viser bl.a. store utvidelser. Dessuten ble malmen fulgt ca. 70 m lengre mot nord. På 280m-etasjen er oppfaringen begynt og malmen fortsetter her i samme kvalitet og mektighet.

Horizontal projection of the three levels of Bleikvassli mine. Black: compact ore (mineable impregnations not included). Dot-dashed line: limitations of the exhausted parts. The map shows the situation in november 1958. In the meantime new ore was found on the 330m-level, e.g. around the thickest parts on the map. Besides, the ore has been followed 70 m farther to the north. On the 280 m-level driving has begun and the ore continues with same quality and thickness.

Joma ligger i et synklinalformet grønnstensparti hvis utgående har linseform. Forekomsten ligger nokså nøyaktig der hvor grønnstens-«puten» er tykkest. De geofysiske anomalier tyder på en stjert-om-stjert-anordning av malmstokkene, med de største mektighetene i midten.

Skorovass-forekomsten er beliggende i et større grønnstensareal som samtidig danner en litt merkelig tverrstruktur (tverrsynklinal ?). Man kunne tenke på en øket mektighet av grønnsten. Av det i BVL's jubileumsbok gjengivne lengdesnitt (ENGZELIUS 1957) kan man se liknende utkilinger som nevnt fra nordre Gjetryggen gruve i Folldal. De som går mot hengen går i motsatt retning av dem som går mot liggen. Det samme fremgår av GJELSVIKS avhandling (1960).

Gjersvik-forekomsten ligger vest for Limingen på et sted hvor grønnstenen etter innsnevringen vest for Namsvatnet har større areal, noe som kanskje tyder på større mektigheter.

Et annet område har jeg allerede nevnt før: Birtavarre hvor VOKES (1957) fant en viss «structural control» av malmen ved at mineraliseringen her er knyttet til et bestemt sted innen Kåfjordsynklinalen. Bortsett fra en forekomst ligger alle der hvor Ankerliaskiferens utgående er bredest. En stjert-om-stjert-anordning av malmplatene nevner VOKES som en mulighet fra Moskogaissa 115 (s. 30). Men forholdene er der ikke så karakteristiske som andre steder og man kan heller ikke vente det, fordi forekomstene er meget små.

Ellers er det et faktum at kismalmen svært sjelden danner enkle plater, og stjert-om-stjert-anordning har fra en rekke forekomster vært kjent lenge. Jeg skal her nevne fra FOSLIE (1926): Kjøli, Heimtjønnhø, Bossmo, Bjørkaasen; fra AASGAARDS arbeide (1927) om kisforekomstene i Tydal og Gauldalen: Storvoll. Han har også til en viss grad vært klar over at malmen sender utkilinger mot hengen og liggen. Det fremgår bl.a. av FOSLIES kart over Kongens grube på Røros samt hans bemerkning at malmen «ofte sender utløpere inn i de liggende skifre» (FOSLIE 1926). Også enkelte ting på AASGAARDS lengde- og tverrprofiler fra Killingdal kan tyde på det samme. CARSTENS (1935) kaller dem «primære kisapofyser» og nevner dessuten at kislegemer på forskjellige skiferhorisonter kan forene seg.

Konklusjon

I det foranstående er noen norske kisforekomster sammenliknet. Det viser seg at disse har visse trekk til felles. Man ser at det eksisterer en viss forbindelse mellom malm og store bergartsmektigheter. For Visnes, Bleikvassli og Joma gjelder det på den måten at de store anrikningene i det hele tatt ligger innenfor de største bergartsmektigheter. Dette gjelder såvidt man nå kan se også for Hjerkinn, men dette kan ikke sies med den samme sikkerhet. Malmen består i disse tilfeller gjerne av ikke bare én skive eller stokk, men av en rekke av dem, hvorav den største danner et slags sentrum, de øvrige er mindre og ligger stjert om stjert forskjøvet mot hengen. Som oftest er denne stjert-om-stjert-anordning mye mer utpreget på den ene siden av «sentret» enn på den andre (f.eks. gamle Visnes, Joma, Bleikvassli). De geofysiske anomalienes regionale anordning — som vanligvis skyldes andre ting enn malm — viser det samme bilde.

For Folldal gjelder alt dette på en annen måte: Her opptrer kisen ikke bare i forbindelse med de største bergartsmektigheter, men der er kobbergehalten også høyest ved de største mektigheter. Tilsvarende er også det geofysiske bilde litt forskjellig: Man ser lange drag med sterke anomalier og der hvor det er uregelmessigheter — som f.eks. brudd i de lange drag, eller en rekke paralleller — der opptrer malmen. (Det er forresten en gammel erfaring av geofysikerne at lange anomalidrag pleier bestå av grafittskifer eller impregnasjonssoner). Men også i Folldal finner man stjert-om-stjert-oppbygning av malmene — eller utkilinger mot hengen og liggen, men disse er jo i grunnen bare en variasjon av stjert-om-stjert-oppbygningen, som Bleikvassli viser.

Til slutt vil jeg forsøke å gi en tolkning av de forannevnte resultater. Som utgangspunkt er det da naturlig å gi en forklaring på fortykkelsen av bergartene som ble nevnt fra Karmøy, Hjerkinn—Folldal, Bleikvassli, Joma og Birtavarre. Det kan ikke dreie seg om en massetransport under foldningen hvor mektighetene gjerne blir større i foldeombøyningene som det er beskrevet bl.a. av BUGGE (1948) fra Dunderlandsdalen. Disse fortykkelsene faller nemlig bare i Joma- og Birtavarrefeltet sammen med synklinalakser. På Karmøy ligger fortykkelsen ved siden av synklinalen og i de to tilfeller Folldal og Bleikvassli ligger fortykkelsene tvers eller på skrå på synklinalen. Dessuten gjelder det her kilometer-dimensjoner, mens det i Dunderlandsfeltet er av størrelsesorden 100 m og med en helt annen intensitet av foldningen. Jeg antar av den grunn at det dreier seg om primære mektighetsvariasjoner som skyldes lokale avsenkningstendenser på havbunnen i sedimentasjons- og erupsjonstiden.

Er det så mulig at kisforekomstenes stjert-om-stjert-oppbygning skyldes et system av parallelle sprekker som er fyllt igjen med malmmineralene? Slikt er jo kjent f.eks. fra jernspatgangene i Siegerland. Når to krefter virker mot hverandre, kan et slikt sprekkesystem oppstå (Fiederklüfte). Disse sprekkesystemene pleier være meget regelmessige og pleier å danne en nokså stor vinkel med strøkretningen. Ingen av delene er tilfelle ved de norske kisforekomstene jeg kjenner. For Jomas og Bleikvasslis vedkommende er det sogar helt umulig å snakke om ett slikt sprekkesystem, her måtte man i tilfelle anta to systemer som ikke lar seg forene med hverandre.

Som før nevnt er kisforekomstene knyttet til store bergartsmektigheter. Forklaringen på kisforekomstenes opptreden må da også søkes i forbindelse med de omtalte — trauaktige — forsenkninger på havbunnen i geosynklinaltiden. Jeg har i en tidligere publikasjon allerede forklart min oppfatning om dannelsen av de norske kisforekomster (Geis 1958). Jeg regner med hydrotermale oppløsninger som er kommet opp i havet og siden blitt utfelt i trauaktige forsenkninger. De i denne avhandling meddelte resultater tillater å bygge ut disse teoriene på følgende måte: Samtidig med bunnfellingen av sulfidene i de langstrakte trauene fortsatte den vanlige sedimentasjon og den vulkanske aktivitet. Herved fyltes trauets bunn etterhvert igjen. Dette skjedde ikke jevnt, men antakelig gikk dette fortere på den siden som lå nærmest opphavsstedet for det sedimentære, vulkanske og eventuelt hydrotermale materiale. Dermed forskjøv lengdeaksen av traubunnen seg etterhvert i en bestemt retning (vekk fra materialtilførselen), og som følge av dette må også bunnfellingsstedet for de hydrotermale oppløsninger ha flyttet seg i tidens løp. Foregikk det en jevn tilførsel av slike oppløsninger, så fikk man en malmplate med utkilinger (f.eks. Nordre Gjetryggen i Folldal), var tilførslen mer ujevn fikk man isolerte malmlinser (f.e. Joma). Hvor det opptrer flere forekomster innenfor ett og samme trau — som i Folldal-området — må det ha vært «hyller» på skråningen med egne trau. Der hvor flere forekomster ligger rett over hverandre (Karmøy, nordre og søndre Gjetryggen) må man anta at avsenkningen foregikk over et lengre tidsrom.

LITTERATUR.

BUGGE, C.: Kisene i fjellkjeden. — NGT, 27, s. 97—102. Oslo 1949.

BUGGE, J. A. W.: Rana Gruber. — NGU 171. Oslo 1948.

CARSTENS, C. W.: Zur Frage der Genesis der norwegischen Kiesvorkommen. — Z. prakt. Geol. 40, S. 97—99. Halle 1932.

— Zur Genesis der Kiesvorkommen des Trondheimgebiets. — NTH.-avhandlinger til 25-årsjubileet 1935. Trondheim 1935.

DITTMARSCH-FLOCON: Über die geologischen und mineralogischen Verhältnisse von Vigsnäs auf Karmöe in Norwegen. — Abh. Sitzungsber. Isis in Dresden 1875.

DU DRESNAY, R.: Quelques observations dans le district Kongsvoll—Snøhetta. — NGT 28, 1950.

ENGZELIUS, J. G.: Elektrokemisk A/S. Skorovass Gruber. — Bergverkenes Landssammenslutning gjennom 50 år 1907—1957. — Oslo 1957.

FALKENBERG, O.: Geologisch-petrografische Beschreibung einiger südnorwegischer Schwefelkiesvorkommen mit besonderer Berücksichtigung ihrer Genesis. — Z. prakt. Geol. 22, S. 105—154. Berlin 1914.

FOSLIE, S.: Norges svovelkisforekomster. — NGU 127. Oslo 1926.

— Kisdistriktet Varaldsøy—Ølve i Hardanger og bergverksdriftens historie. — NGU 147. Oslo 1955.

— Geologisk kart Namsvatnet med en del av Frøyningsfjell 1:100000. — NGU 196. Oslo 1957.

— Geologisk kart Trones 1:100000. Oslo 1958 (1958a).

— Geologisk kart Tunnsjø 1:100000. Oslo 1958 (1958 b).

— Geologisk kart Sanddøla 1:100000. Oslo 1958 (1958 c).

GEIS, H.-P.: Die Genese der norwegischen Kieslagerstätten. — Erzmetall 11, S.541—543. Stuttgart 1958.

GJELSVIK, T.: The Skorovass pyrite deposit, Grong area, Norway. — Int. Geol. Congr., Rep. 21. session, part 16, p. 54—66. Copenhagen 1960.

GOLDSCHMIDT, V. M.: Geologisch-petrografische Studien im Hochgebirge des südlichen Norwegens. V. Die Injektionsmetamorphose im Stavanger—Gebiete. — Vid. selsk.skr. I. Mat.nat. Klasse 1920, no. 10. Kristiania 1921.

HELLAND, A.: Ertsforekomster i Søndhordland og paa Karmøen. — Nyt Mag. f. Nat., 18. Bd., S. 225—275. Christiania 1871.

HERNES, I.: Geologisk oversikt over Molde—Kristiansundsområdet. -- Kgl. Norske Vidensk. Selsk. Skr. 1955, nr. 5. Trondheim 1956.

HJELSETH, G. & EINARSEN, S.: Folldal Verk. — Tidsskr. f. kjemi, bergvesen og metall, nr. 2 b, s. 70—73. Oslo 1950.

— Folldal Verk. — Bergverkenes Landssammenslutning gjennom 50 år 1907—1957, s. 129—138. Oslo 1957.

HOLMSEN, P.: Trekk av Oppdalsfeltets geologi. — NGT 35. Bergen 1955.

ISACHSEN, F.: Fossilførende sedimenter ved Skudenes. — NGU 155, s. 5—18. Oslo 1940.

JONSSON, I.: Gruvedriften ved A/S Stordø Kisgruber. — Tidsskr. f. kjemi, bergvesen og metall, 9, s. 149—156. Oslo 1952.

KNUDSEN, E.: Nogle Bemærkninger om Ertsforekomstene ved Vigsnæs Grube. — Nyt. Mag. f. Nat., 29 bd., s. 306. Christiania 1885.

KOLDERUP, N.-H.: Oversikt over den kaledonske fjellkjede på Vestlandet. — Bergens Museums Årbok 1931, nat.-vid. rekke nr. 1. Bergen 1931.

KVALE, A.: Et kaledonisk intrusiv- og effusivfelt på Stord. — Bergens Museums Årbok 1937, nat. vid. rekke no. 1. Bergen 1937.

MARLOW, W.: Folldal. Beskrivelse til det geologiske rektangelkart. — NGU 145. Oslo 1935.

OFTEDAHL, C.: Oversikt over Grongfeltets skjerp og malmforekomster. — NGU 202. Oslo 1958 (1958 a).

— A Theory of Exhalative-Sedimentary Ores. — GFF 80, H. 1, s. 1—19. Stockholm 1958 (1958 b).

REKSTAD, J.: Hatfjelldalen. Beskrivelse til det geologiske generalkart. — NGU 124. Oslo 1924.

REUSCH, H.: Fjeldbygningen ved Viksnes Kobbergrube på Karmøen. — Nyt. Mag. f. Nat., 28. Bd. Christiania 1884.

— Bømmeløen og Karmøen med omgivelser. — Kristiania 1888.

SKORDAL, A. J.: Vulkanitter og sedimenter på den sørøstre del av Stord. — Univ. i Bergen Årbok 1948, natvid. rekke no. 2. Bergen 1950.

VOGT, J. H. L.: Norske ertsforekomster. III. række. VIII. Foldalens kisfelt. — Ark. f. mat. og nat., 13. bd. Kristiania 1889.

— Über die Kieslagerstätten vom Typus Röros, Vigsnäs, Sulitjelma in Norwegen und Rammelsberg in Deutschland. — Z. prakt. Geol. 12, S.41—50, 117—134 & 173—181. Halle 1894.

— Om relationen mellem størrelsen av eruptivfelterne og størrelsen av de i eller ved samme opptrædende malmutsondringer. — NGU 43, III. Kristiania 1905.

— I Beyschlag—Krusch—Vogt: Die Lagerstätten der nutzbaren Mineralien und Gesteine. I. Bd. Stuttgart 1910.

VOGT, T.: Fjellkjedens flytestrukturer og malmforekomster. I. Nord-Rana grubefelt. — Det kgl. Norske Vid.Selsk.Forh., Bd. XVII, nr. 30. Trondheim 1945.

— Flowage structures and ore deposits of the Caledonides of Norway. — Int. Geol. Congr., Rep. of 18th Session, part XIII, p. 240—244. London 1948.

— A Lateral Crustal Movement in the Caledonids of Norway. — Kgl. Norske Vid. Selsk. Forh. 26, nr. 23, s. 102—109. Trondheim 1954.

VOKES, F. M.: The Copper Deposits of the Birtavarre District, Troms, Northern Norway. — NGU 199. Oslo 1957.

AASGAARD, G.: Gruber og skjerp i kisdraget øvre Guldal—Tydal. — NGU 129. Oslo 1927.

GEOFYSISK MALMLETING: A/S Folldal Verk. Elektrisk undersøkelse Nordre grube. Folldal 17.4.—17.5.41. Trondheim 1941.

GEOFYSISK MALMLETINGS rapport nr. 103. Geofysiske undersøkelser Tverfjellfeltet. Hjerkinn, Dovre hd. Trondheim 1953.

GEOFYSISK MALMLETING: Elektromagnetiske undersøkelser vest for Hovedgruben, Folldal. Kart 1:10000. Trondheim 1957.

Manuskript mottatt 10. desember 1960.

Trykt desember 1961.

THE SCANDIUM CONTENT OF SOME NORWEGIAN MINERALS AND THE FORMATION OF THORTVEITITE, A RECONNAISSANCE SURVEY

By

HENRICH NEUMANN

In their classical paper on the geochemistry of scandium V. M. GOLDSCHMIDT and CL. PETERS (1931) present a great deal of information about the scandium content of Norwegian minerals of varied geological origin. Also G. EBERHARD (1908 and 1910) used, among others, samples from this country for his investigations which demonstrated the common and widespread occurrence in the earth's crust of the element which had until then been regarded as very rare and scarce. Further data are given by I. OFTEDAL (1943), who was especially interested in the variations of the scandium content of biotite as a function of temperature, and the possibility of using this as a geological thermometer.

Thortveitite, the only known, and very rare, primary mineral of scandium occurs in some quantity in several pegmatite dykes in the Iveland—Evje district North of Kristiansand.[1]

---

[1] Thortveitite has been found in the following quarries: In Iveland: Steane, Frikstad; Tuftane (= Frikstad No. 3); Mannekleiv (= Håverstad No. 1); Eptevann No. 2; Knipan (first find of thortveitite, = Ljosland No. 4; and also a recently opened quarry about 30 m from the old one); Knapen, Birkeland (opened a few years ago); Slobrekka (= Frikstad No. 7); Frikstad school (about 150 m North of Frikstad schoolhouse, on a small hill between the main road and the road to Frikstad); Ivedal (= Ivedal No. 1); Torvelona (= Ljosland No. 14); Eretveit (= Ertveit No. 1); Birkelandjordet (just North of

(cont., p. 198)

The reason for this is obscure. It may be recalled that V. M. Goldschmidt (1934) suggested that the scandium content of thortveitite may not originate from the pegmatite magma itself as scandium is not a characteristic element of acidic rocks; on the contrary it is enriched in basic rocks. The thortveitite-bearing pegmatites are always found in amphibolite, and he found it more reasonable to think that the pegmatite magma has extracted scandium from the surrounding amphibolites with a subsequent crystallization of thortveitite as a normal event during the later freezing of that magma. It is an interesting fact that the bed rock of the greater part of the very numerous pegmatites in the precambrian of Southern Norway is amphibolite (Barth (1928)). If the theory outlined above is correct one would expect a higher content of scandium in the amphibolites surrounding the thortveitite-bearing pegmatites than in the amphibolites surrounding pegmatites without thortveitite. The data are too scanty for any very definite statement to be made. The few data available give, however, no indication of such a difference (Oftedal (1943) p. 205). An alternative explanation would be to postulate an unexplained property of the „magma'' or pegmatizing medium of a thortveitite-bearing dyke, which enables it to extract scandium more efficiently from the surrounding amphibolite than is the case with the pegmatizing medium of a pegmatite that does not carry thortveitite. Such a working hypothesis may be perfectly sensible, but would seem to be without any sound basis in observed facts. The present author has felt for some time that Goldschmidt's hypothesis is not substantiated by the knowledge about the pegmatites in the Iveland—Evje district, which has accumulated since it was offered for consideration.

(cont. from 197)

the Birkeland field East of the road, 100 m from Birkeland No. 1); Tjomstøl; In Evje: Landsverk No. 3 (= Beryl-bruddet = Hagebruddet); Unneland No. 2 (= Beryl-bruddet); and in Vegusdal: Vatne.

According to Mr. Olaf Landsverk, Iveland, the following approximate quantities of thortveitite have been produced from the several quarries: Steane 15—20 kgs, Tuftane 12—15 kgs, Mannekleiv 5—6 kgs, Eptevann 3—4 kgs, Knipan 3—4 kgs, Knapen 1 kg, Slobrekka 1/2 kg, Frikstad school 1/2 kg, Landsverk about 6 kgs, Unneland 1 kg, and from the others less than 1/2 kg. Altogether, the Norwegian production of thortveitite amounts to about 50 kgs, corresponding to not much more than 20 kgs of scandium metal.

As an attempt to get somewhat closer to the solution of the genetical problem of the thortveitite-bearing dykes, it seemed highly desirable to aquire further data on the scandium content of pegmatite minerals from the Iveland—Evje district as well as from elsewhere. The scandium determinations were undertaken by Professor Ivar Oftedal with usual generosity. I should like to take this opportunity to express my sincere thanks for all the help he has given me with these investigations as well as in other connections. The figures as presented in Table 1 are only semiquantitative in this sense of the word that the true values are supposedly between 2/3 and 3/2 of the given figures. The reader is referred to OFTEDAL's (1943) paper on scandium in biotites for an evaluation of the correctness of the method and for further details.

At the request of J. Schetelig a few spectrographic examinations were made by G. Eberhard of the content of scandium in minerals from four thortveitite-bearing dykes. An appreciable content of scandium in beryl is indicated (J. SCHETELIG (1922) pp. 84—85), and OFTEDAL (1943, p. 206) reports a content of about 10,000 ppm (1% Sc) in an ordinary green beryl from Ljoslandsknipan. As beryl is a fairly common mineral in Iveland, a content of scandium of that order of magnitude would even be of considerable practical importance and make such beryls a valuable source of scandium. Scandium determinations were therefore made of 57 beryls from Iveland and Evje and a number of other districts. Because of the similarity in crystal structure cordierites of varied geological histories from 9 different localities were also examined. Results are given in Table 1 and figure 1, b and c. Cordierites are low in scandium with approximately 10—20 p.p.m. Sc, while the analyzed beryls vary in scandium content from approximately 10 p.p.m. to 1000 p.p.m. (0.1% Sc).

The data reproduced in figure 1, b indicate very clearly that there is a tendency for beryls from thortveitite-bearing dykes to be richer in Sc than beryls from dykes with no thortveitite, as might be expected. It is a complication in this connection that several varieties of beryl may occur in one individual dyke, the beryls apparently belonging to different generations and varying stages of the formation of the pegmatite. If a list is extracted from Table 1 of beryls with less than 100 p.p.m. Sc from thortveitite-bearing dykes, it reads as follows:

Aquamarine	Mannekleiv, Håverstad, Iveland	20	p.p.m.
Aquamarine	„The beryl quarry", Landsverk, Evje	20	—
Aquamarine	Landsverk 3, Evje	20	—
Green beryl	—	50	—
Altered beryl	—	50	—

It is hardly a coincidence that not less than 3 of these 5 beryls are aquamarines, which are supposedly amongst the very youngest minerals of the dykes concerned. The aquamarine from „The thortveitite quarry" at Eptevand, however, has a scandium content of 300 p.p.m. The „normal" beryl from Mannekleiv, Håverstad, contains 500 p.p.m. scandium as compared to the 20 p.p.m. of the aquamarine from the same deposit. Of the three scandium-rich beryls from outside Iveland and Evje one is from Hitterø near Flekkefjord, and the two others from a nordmarkite pegmatite dyke at Byrud in Minnesund (V. M. Goldschmidt (1911) p. 56). The high scandium content, especially of the Byrud beryls, is rather surprising and would seem to warrant further investigations.

The fact that the minerals from thortveitite-bearing dykes are in general richer in scandium than the same minerals from dykes without thortveitite is convincingly demonstrated by the data for biotites given by Oftedal (1943). His data are reproduced in this paper as figure 1,a. Unfortunately 4 biotites only from thortveitite-bearing dykes have been analyzed, they contain from 500 p.p.m. to 1000 p.p.m. Sc, while all the 17 biotites from dykes without thortveitite are poorer in Sc with an average Sc content of less than 100 p.p.m.

Because of the similarity in ionic radii a high content of scandium might be expected in zirconium minerals. It is indicated by data published earlier that that is not so, and this is confirmed by further data presented in this paper, see Table 1 and figure 1, d. All the zirconium silicates which were analyzed contain very little scandium, approximately 10 p.p.m., except 2 zircons, which are both from the island of Seiland in Finnmark, one from a canadite pegmatite, and the other which is richer in Sc from a plumasite pegmatite (see Barth (1927)). The identical location of the two zircons is hardly altogether accidental, but no comment can be offered at present as no other minerals from pegmatites on Seiland have been analyzed for scandium, nor is any other information available which would seem to be relevant.

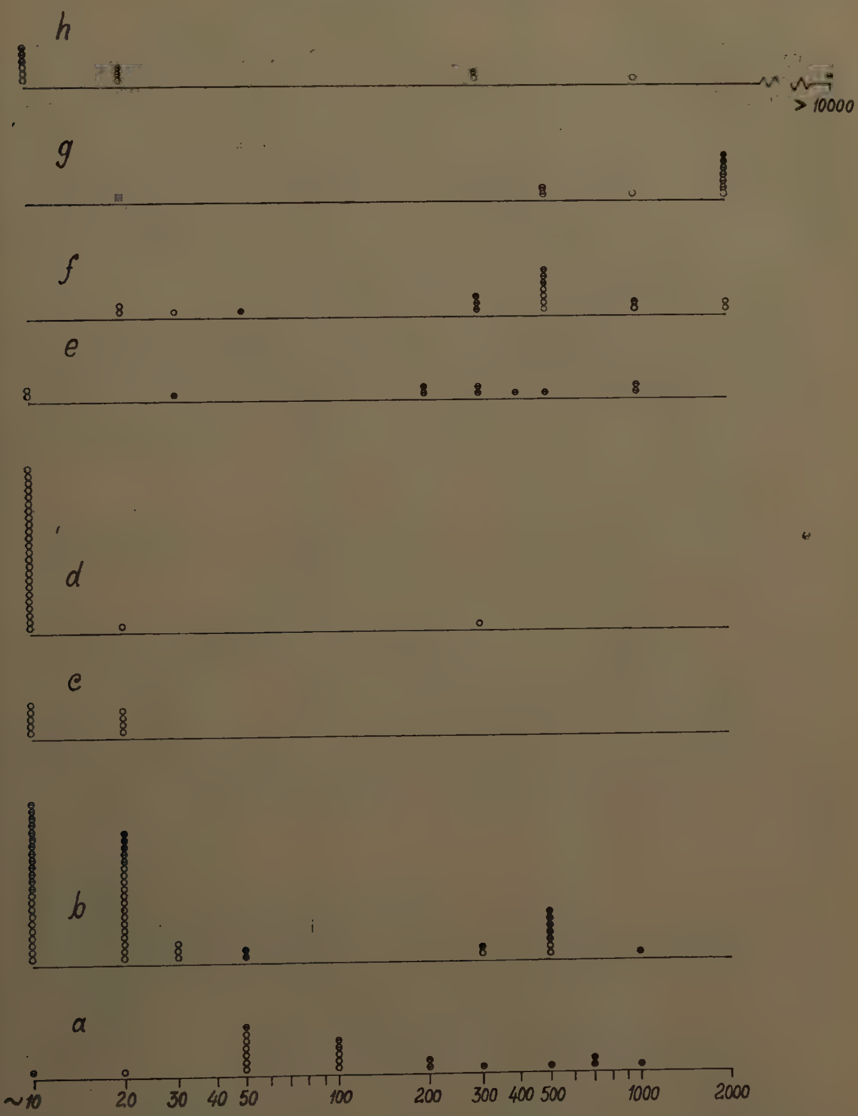


Fig. 1. Sc in p.p.m. in *a*: biotite (data from I. OFTEDAL (1943)), b: beryl, *c*: cordierite, *d:* zirconium silicates, *e:* ilmenite, *f:* columbite, *g:* ilmenorutile, and *h:* xenotine.

Filled circles: thortveitite-bearing deposits in Iveland and Evje. Circles with lines: deposits in Iveland and Evje which are not known to contain thortveitite. Open circles: deposits elsewhere in Norway.

Earlier publications give very limited information about the scandium content of ilmenites and columbites from pegmatite dykes, in fact only one scandium determination is available of a columbite from Ånnerud near Moss in Norway, containing 50 p.p.m. Sc_2O_3. Analyses of 11 ilmenites and 18 columbites are given in Table 1 and figure 1, e and f. They prove a significantly high content of scandium in these minerals with an average of 585 p.p.m. Sc in columbites and 435 p.p.m. Sc in ilmenites from pegmatite dykes.

Ilmenorutiles are higher in scandium than any other mineral investigated so far (see figure 1, g) with an average content of 1455 p.p.m. Sc. Also in the case of this mineral there is a clear tendency for a higher content of scandium in thortveitite-bearing dykes than in dykes without thortveitite, even if that tendency is not as pronounced as is the case with biotite.

Because of the fairly common occurrence of phosphates of the rare earth elements it seemed worth while to look for the chemically related scandium in pegmatite phosphate minerals.

With one exception the examined apatites are all very poor in scandium, and so are the wagnerites from the Bamble apatite deposits in spite of the similarity in ionic radii of Mg and Sc. In this connection it is worthy of notice that the one apatite with a higher content of scandium (500 p.p.m. Sc) is from the Ödegården apatite deposit in Bamble (W. C. Brögger and H. H. Reusch (1875)) while three analyzed wagnerites from the same deposit contain 10 p.p.m. Sc only.

Xenotimes are on the whole fairly poor in scandium (see figure 1, h), except one xenotime from the Omland pegmatite quarry in Evje which contains more than 1 percent Sc. The scandium content is not sufficiently high to influence the size of the unit cell of that mineral as shown by an X-ray powder diagram.

The few davidites from Iveland which were analyzed contain 200—500 p.p.m. Sc.

Magnetites do not seem to be appreciably enriched in scandium, even in the case of a magnetite from a thortveitite-bearing dyke (Ivedal, Iveland). The sample of magnetite from Mölland with 300 p.p.m. Sc was contaminated by ilmenite.

The aegirite (acmite) from the alkaline granite pegmatite (ekerite pegmatite) at Rundemyr, Eiker, is surprisingly low in scandium in agreement with data published earlier by GOLDSCHMIDT and PETERS (1931, p. 265).

As mentioned above there is a clear tendency for minerals from thortveitite-bearing dykes in Iveland and Evje to be richer in scandium than the same minerals from dykes that do not carry thortveitite. In this connection it would seem of interest to examine if there is any difference in the scandium content of minerals from dykes without thortveitite from the Iveland—Evje district as compared to dykes (without thortveitite) from other districts. It appears from the following list that there is no obvious trend as the figures for biotite and ilmenorutile on the one side and beryl and columbite on the other are contradictory.

	Iveland—Evje district	Other districts
Biotite	135 p.p.m. Sc	60 p.p.m. Sc
Beryl	40 —	60 —
Columbite	430 —	710 —
Ilmenorutile	1500 —	(1005) —

(Figures cannot be given for ilmenite as the two samples from outside Iveland and Evje, which are poor in scandium are not from pegmatime dykes. The xenotime figures would be misleading because of the exceptionally high scandium content of one sample, see figure 1, h.) Data are too few, but it may be tentatively concluded that there is no systematic difference in the scandium content of minerals from dykes without thortveitite in the Iveland—Evje district and from dykes in other districts.

In conclusion the relevant features which are apparent from the data presented in this paper and earlier data may be summarized as follows:

1. The bulk of the scandium content of granite pegmatites is contained in their iron minerals in complete agreement with the long established fact that the diadochy of scandium and divalent iron is the dominating factor in the geochemistry of scandium. Most important are the common pegmatite minerals biotite and ilmenite, while other minerals, as for example, columbite and ilmenorutile may have a higher content of that element. Allanite has not been analyzed, but may be of considerable interest in this connection.

2. Minerals from dykes with thortveitite are richer in scandium than the same minerals from dykes without thortveitite.

3. In the dykes that do not contain thortveitite there does not appear to be any systematic variation in the scandium content of the pegmatite minerals from one part of the country to another. Minerals from such dykes in „the thortveitite district" of Iveland and Evje are not any richer in scandium than minerals from dykes in other districts.

4. It is indicated that the amphibolitic bed-rocks of the thortveitite-bearing dykes are not any richer in scandium than the amphibolitic bed-rocks of other dykes within or outside the Iveland—Evje district. More data are needed before this question can be definitely settled.

The content of thortveitite in certain dykes in the Iveland—Evje district, and the observed facts about these dykes are easily explained if it is assumed that the thortveitite-bearing pegmatites are poorer in divalent iron than normal. In the vast majority of pegmatite bodies the entire scandium content is found in the iron minerals. If, as a hypothetical experiment, the content of divalent iron in a pegmatite is decreased while the content of scandium is kept constant the content of scandium in the ever smaller amount of iron minerals will increase until a stage is reached when these minerals are saturated in scandium under the prevailing P—T—X-conditions. With a further decrease in iron, scandium will necessarily have to form a mineral or minerals of its own as the limit of solid solution is reached. If this hypothesis is correct, the thortveitite-bearing dykes should have less iron minerals like biotite and ilmenite than other dykes in the district with apparently closely similar geological histories. This may be so, but there are no data available to prove it, in fact it is the very crux of pegmatite research that it is in general well-nigh impossible to get reliable quantitative figures for the relative amounts of the pegmatite-forming minerals or for the chemical composition of the entire pegmatite mass. There is, however, a piece of circumstantial evidence for a deficiency in iron of the dykes concerned. It has been known for a long time that ilmenorutile[1] is a characteristic mineral of thortveitite-bearing dykes while it is a pretty rare mineral in other dykes of the district. With an excess of iron ilmenite is obviously the stable titanium mineral as demonstrated by its very common occurrence in pegmatites, and there is hardly much doubt that rutile or ilmenorutile can safely be regarded as indicators of a deficiency in divalent iron in a pegmatite containing these minerals. It is also worthy of notice that of the 15 thortveitite-bearing dykes in Iveland and Evje BJÖRLYKKE (1934) reports ilmenite from only two of them. Because of the generally high content of scandium in ilmenites, it would be of great value to have even semi-quantitative or subjectively evaluated data for the relative amounts of ilmenite and biotite in dykes with thortveitite as contrasted to dykes without thortveitite. Figures for the iron content of the garnets of those dykes would be of considerable interest, and the assumed deficiency in iron may even be reflected in the chemical composition of the micas. Unfortunately, no such data are available at present.

---

[1] In the case of the Tuftan pegmatite rutile instead of ilmenorutile occurs together with thortveitite.

Table 1. Scandium content in p.p.m. Sc of 176 mineral samples.

Mineral	Locality	Sc in p.p.m.
Beryl:		
	(Emerald) Byrud. Minnesund. Eidsvoll	500
	Skarpnes. Øiestad. Aust-Agder	20
	St. Antoinette's mine. Straumsfjord	20
	Brudalen. Kviteseid. Telemark	20
	Aspedammen. Idd. Østfold	20
	Aspedammen. Idd. Østfold	20
	1050 m N. of Rørvik molybdenite mine. Hurum	30
	(Green crystal). Ånnerud. Near Moss	20
	(Colourless crystal). Ånnerud. Near Moss	20
	Halvorsrud. Råde. Østfold	10
	Buvasshei. Nisserdal. Telemark	20
	(Yellow crystal). Elvestad. Råde. Østfold	20
	(Light red crystal). Høydalen seter. Tørdal	20
	(Inner part of large crystal). Southern Ski. Rakkestad. Østfold	20
	Berby. Østfold	20
	Ødegårdsletten. Våler. Østfold	20
	Halvorsrud. Råde. Østfold	20
	(Emerald). Byrud. Minnesund. Eidsvoll	300
	Årvoldskåven. Vansjø. Near Moss. Østfold	30
	(Yellow crystal). Håvåsen. Iveland	10
	Halvorsrud. Råde. Østfold	10
	Ånnerud. Near Moss	20
	Aker. Råde. Østfold	10

Mineral	Locality	Sc in p.p.m,
	Vatvedt. Rakkestad. Østfold	10
	Boksjøen mineral mine. Idd. Østfold	10
	Southern quarry. Vådne. Iveland	10
	Frikstad. Iveland	10
	Eitland. Austa. Vest-Agder	10
	(Yellow, clear crystal). Støledalen. Iveland	10
	Bjertnes feldspar quarry. Krødsherad. Buskerud.	10
	«The beryl quarry». Evje	20
	(Aquamarine). «The thortveitite quarry». Eptevann. Iveland	300
	(Aquamarine). Mannekleiv. Håverstad. Iveland	20
	Hitterø. Near Flekkefjord	500
	"The thortveitite dyke". Ljoslandsknipa. Iveland	1000
	Heia. Håverstad. Iveland	10
	"The thortveitite quarry". Steane. Nedre Frikstad. Iveland	500
	(Aquamarine). "The beryl quarry"..Landsverk. Evje	20
	8. Dalane. 2. Iveland	10
	(With topaz). By the tarn, Nedre Frikstad. Iveland	10
	(Yellow crystal). Steli. 3. Tveit. 3. Iveland	10
	(Yellow crystal). 54. Mølland. 3. Iveland	10
	Tveit. Iveland	10
	(Blue crystal). Mannekleiv. Håverstad. Iveland	500
	Tveit. Iveland	10
	"The beryl quarry". Håverstad. Iveland	500
	(Piece of a crystal weighing 2200 kgs). Håvåsen. Eptevann. Iveland	10
	(Large crystal). Rona. Randøysundfjord. Near Kristiansand. Vest-Agder	10
	Ollestad quarry. Heskestad. Rogaland	10
	Ljosland. Knipan. Iveland	500
	(Aquamarine). Håvåsen. Eptevann. Iveland	10
	Rona. Randøysundfjord. Near Kristiansand. Vest-Agder	10
	Stetind feldspar quarry. Tysfjord. Nordland	30
	Knipan 2. 30 m from Knipan 1. Iveland	500
	(Aquamarine). Landsverk 3. Evje	20
	Landsverk 3. Evje	50
	(Altered crystal). Landsverk 3. Evje	50
Cordierite:		
	Rød station. Risør	10
	Kammerfors. Kragerø	10
	Bromsbo mine. Tvedestrand	10
	Frøina. Søndeled.	20
	Sandøkedal	10
	Bjordammen. Bamble	20

Mineral	Locality	Sc in p.p.m.
	Lindland. Søndeled	20
	Risø feldspar quarry. Kragerø	20
	Klovstenen. Østebø. Søndeled	10
Zircon:		
	Syenite pegmatite east of Herregårdsdammen. Larvik	10
	Stavern. (= Fredriksvern). Near Larvik	10
	Østerholtheia. Gjerstad	10
	Barkevik. Langesundsfjord	10
	Southern precipice of Stokø. Langesundsfjord	10
	Mørjefjord. East of Langesundsfjord	10
	Small canadite pegmatite on the farm-yard of Southern Skarvbergnes. Seiland. Finnmark	20
	Tveitdalen	10
	Plumasite pegmatite. SW of Skarvvann. Seiland. Finnmark	300
Eucolite:		
	Arøsund. Langesundsfjord	10
	Arøskjær. Langesundsfjord	10
	Brattholmen. Langesundsfjord	10
	Sandø. Langesundsfjord	10
Hiortdahlite:		
	Southern part of Store Arø. Langesundsfjord	10
Eudialyte:		
	Brattholmen. Langesundsfjord	10
	Langesundsfjord	10
Catapleiite:		
	Arø. Langesundsfjord	10
	Låven. Langesundsfjord	10
	Eikaholmen. Langesundsfjord	10
Rosenbuschite:		
	Låven. Langesundsfjord	10
Mosandrite:		
	Låven. Langesundsfjord	10
	Langesundsfjord	10
Johnstrupite:		
	Barkevikskjær. Langesundsfjord	10
	Skudesundskjær. Langesundsfjord	10

Mineral	Locality	Sc in p.p.m.
Wöhlerite:		
	Skudesundskjær. Langesundsfjord	10
	Barkevikskjær. Langesundsfjord	10
Ilmenite:		
	Hilltveit. Iveland	200
	Landsverk. Evje	200
	Tveit. Iveland	30
	South of the marsh Knipan. Ljosland 11. Iveland	1000
	Road metal quarry at Valleveien. Ødegården. Bamble. Telemark	10
	Mølland and Ljosland. Iveland	300
	Highest point of Dillingø near Moss. Østfold	10
	(Tabular). Landsverk. Evje	400
	Bjørnekra. Årsland. Evje	300
	Iveland	500
	(With some hematite). Ljosland. Iveland	1000
Columbite:		
	Øvre Gjerstad	20
	Håverstad. Iveland	500
	Klep. Iveland	500
	Eptevann. Iveland	300
	Tveit. Iveland	1000
	Thor's mine. Øvre Vats. Rogaland	20
	Ljoslandsåsen. Iveland	500
	Stelibruddet. 3 Tveit 3. Iveland	300
	63 Ljosland 1. Ljoslandjordet 1. Iveland	300
	Tangen quarry. Kammerfosselv near Kragerø	30
	Lorebø near Aker. Råde. Østfold	2000
	Skråtorp. Råde. Østfold	500
	Vintergruben. Ånnerud near Moss. Østfold	1000
	Rygge. Near Moss. Østfold	500
	Elvestad. Råde. Østfold	2000
	Karlshus. Råde. Østfold	500
	Halvorsrud. Råde. Østfold	500
	Heia. Håverstad. Iveland	50
Ilmenorutile:		
	Ljosland. Iveland	2000
	Lundekleven. Evje	500
	Tveit. Iveland	500
	"The thortveitite quarry". Eptevann. Iveland	2000
	Tjønstøl. Iveland	2000
	"The thortveitite quarry". Ljoslandsknipa	2000

Mineral	Locality	Sc in p.p.m.
	Tuten 2. Håverstad. Iveland	2000
	Håverstad 1. Iveland	2000
	Monbua. Tromøya. Aust-Agder	2000
	Ausel. Near Tvedestrand. Aust-Agder	1000
	Ramskjær. Søndeled. Aust-Agder	20
Xenotime:		
	Holer. Råde. Østfold	20
	"Eikeråsen". Eretveit. Iveland	10
	Omland. Evje.	10000
	Ljosland. Iveland	20
	Høgtveit. Evje	300
	Narestø. Arendal. Aust-Agder	10
	Frikstad. Iveland	10
	Gryting. Gjerstad	1000
	1. Tveit. 1. "Eliasgruva". Iveland	10
	Rasvåg. Hitterø. Near Flekkefjord. Vest-Agder	300
	Smålian. Frikstad. Iveland	20
	Østre Moland. Aust-Agder	10
	Hummerli. Hitterø. Vest-Agder	10
Apatite:		
	Risø. South of Langø. Near Kragerø. Telemark	10
	Husås. Søndeled. Near Risør. Aust-Agder	10
	Skarvstøl. Gjerstad. Aust-Agder	10
	East claim. Holtet mine. Bamble. Telemark	10
	Kjørstad feldspar quarry. Bamble. Telemark	10
	Karlshus. Råde. Østfold	10
	Fennefosheien. Hornnes. Aust-Agder	10
	"The thortveitite quarry". Håverstad. Iveland	10
	Knipan. Ljosland. Iveland	10
	Birkeland. Iveland	10
	Håverstad. Iveland	10
	Frikstad. Iveland	10
	Havredal. Bamble. Telemark	10
	(Red) Ødegården. Bamble. Telemark	10
	(Grey) Ødegården. Bamble. Telemark	10
	(Green) Ødegården. Bamble. Telemark	500
Wagnerite:		
	(Grey) Ødegården. Bamble. Telemark	10
	Ødegården Bamble. Telemark	10
	Ødegården. Bamble. Telemark	10
	Havredal. Bamble. Telemark	20

Mineral	Locality	Sc in p.p,m.
	Havredal. Bamble. Telemark	10
	Havredal. Bamble. Telemark	10
	Skaugen. Bjørdammen. Bamble. Telemark	10
Davidite:		
	Tuftan. Iveland. (Analyzed material)	500
	Birkeland quarry. Iveland	200
	Feldspar quarry in Iveland	500
Magnetite:		
	Ånnerud near Moss. Østfold	10
	Frøyså. Iveland. (With gadolinite)	20
	(Tabular, with ilmenite impurities) Mølland. Iveland	300
	Ivedal. Iveland	20
Acmite:		
	Rundemyr. Eiker. Vestfold	20

REFERENCES.

BARTH, TOM.: Die Pegmatitgänge der kaledonischen Intrusivgesteine im Seiland-Gebiete. Vid.-Akad. Skr. I. M.-N. Kl. 1927. No. 8.

— Zur Genese der Pegmatite im Urgebirge. Neues Jahrb. f. Min. & c., BB. 58A. 1928.

BRÖGGER, W. C. and REUSCH, H. H.: Vorkommen des Apatit in Norwegen. Zeitschr. d. Deutsch. Geol. Gesellschaft 27. 1875.

EBERHARD, G.: Über die weite Verbreitung des Scandium auf der Erde. Sitz. Ber. d. Akad. d. Wissenschaften. Berlin. 1908. II.

— Über die weite Verbreitung des Scandium auf der Erde. II. Ibid. 1910. I

GOLDSCHMIDT, V. M.: Die Kontaktmetamorphose im Kristianiagebiete. Vid.-Selsk. Skrifter. I. M.-N. Kl. 1911. No. 1.

— Drei Vorträge über Geochemie. Geol. Fören. Förh. 56. 1934.

— and PETERS, CL.: Zur Geochemie des Scandiums. Nachr. d. Gesellsch. d. Wissenschaften Göttingen. Math.-Phys. Kl. 1931.

OFTEDAL, IVAR: Scandium in biotite as a geological thermometer. Norsk Geol. Tidsskr. 23. 1943.

SCHETELIG, J. in W. C. BRÖGGER, TH. VOGT, and J. SCHETELIG: Die Mineralien der Südnorwegischen Granitpegmatitgänge II. Vid.-Selsk. Skrifter I. M.-N. Kl. 1922. No. 1.

Manuscript received May 10, 1961.

Printed December 1961.

NORSK GEOLOGISK TIDSSKRIFT 41

A REGIONAL GRAVITY STUDY OVER THE PERMIAN BÆRUM CAULDRON OF THE OSLO REGION

By

S. B. Smithson

Abstract. Thirty-five gravity stations were added to a pre-existing gravity net over the Bærum cauldron and the Bouguer anomalies were computed. A positive anomaly lies over the south side of the cauldron. If the effect of the 1000-m-thick B_3 basalt, which is down faulted into the cauldron, is removed, a 10-mgal. negative anomaly is found over the cauldron. This anomaly corresponds to a hypothetical vertical cylinder of nordmarkite 10 km in diameter and 6 km deep. Its top would be about 3 km below the surface. A small negative anomaly associated with contact-metamorphosed sandstones outside the cauldron is ascribed to a subjacent intrusive. A large negative anomaly is found over the nordmarkitic pluton comprising Tryvasshögda.

Introduction.

In 1960 the Geographical Survey of Norway (Norges Geografiske Oppmåling) published a Bouguer gravity map (1:250 000) over the Oslo region. This map, based on 1400 gravity stations, represents a tremendous amount of work; however, since a large area was covered, many features which are interesting to geologists remain undefined due to an insufficient density of stations. The Bærum cauldron was one of these features and was selected for further study because of its general interest and accessibility. Accordingly, 35 more gravity stations were added over the Bærum cauldron and in its vicinity.

The Bærum cauldron in elliptical in plan and measures 8.5 by 12 km across. The southeastern border of the cauldron lies 7 km northwest of the city of Oslo. The terrain consists of rounded, forest-covered hills whose maximum relief is 300 m.

The purpose of the survey was to determine the gravity anomaly associated with the cauldron and to suggest a course for future measurements in the area. The 35 additional stations hardly provide detailed coverage; consequently, the conclusions can only be proportionately exact.

Field Methods and Reduction of Observations.

A Worden gravimeter with a scale constant of 0.09 mgal. was used. Several local base stations were selected from those already established in the area by the Geographical Survey of Norway, and these were tied in with the base gravity station in the Geologisk Museum, Oslo. Gravimeter drift was corrected by returning the instrument to a local base station twice a day.

The main problem of the gravity survey was to establish accurate elevations at gravity stations. Wherever possible, accurately surveyed elevations were used; otherwise, either spot elevations on lakes and farms from rectangle maps of the Geographical Survey of Norway or barometric elevations were used. Spot elevations on more modern maps are accurate to ± 1 m (O. Trovaag, oral communication). Checks with a Short and Mason barometer between known elevations during a short time interval (one hour or less) showed an accuracy of 2—3 m. Most of the barometric elevations should be within this accuracy; therefore, they may cause an error of 0.6 mgal. in the Bouguer anomalies. Horizontal control was provided by the above-mentioned rectangle maps (1:100 000).

Bouguer anomalies were computed according to NETTLETON, (1940) and a surface density of 2.67 was used for the Bouguer correction. The terrain effect introduces errors up to 3 mgal. at some stations; therefore, terrain corrections, computed using two-dimensional models (HUBBERT, 1942), were applied to most of the stations. These models only approximate the actual terrain but would remove most of the error due to topography. Any error remaining due to terrain effects should not exceed 0.5 mgal. at its greatest.

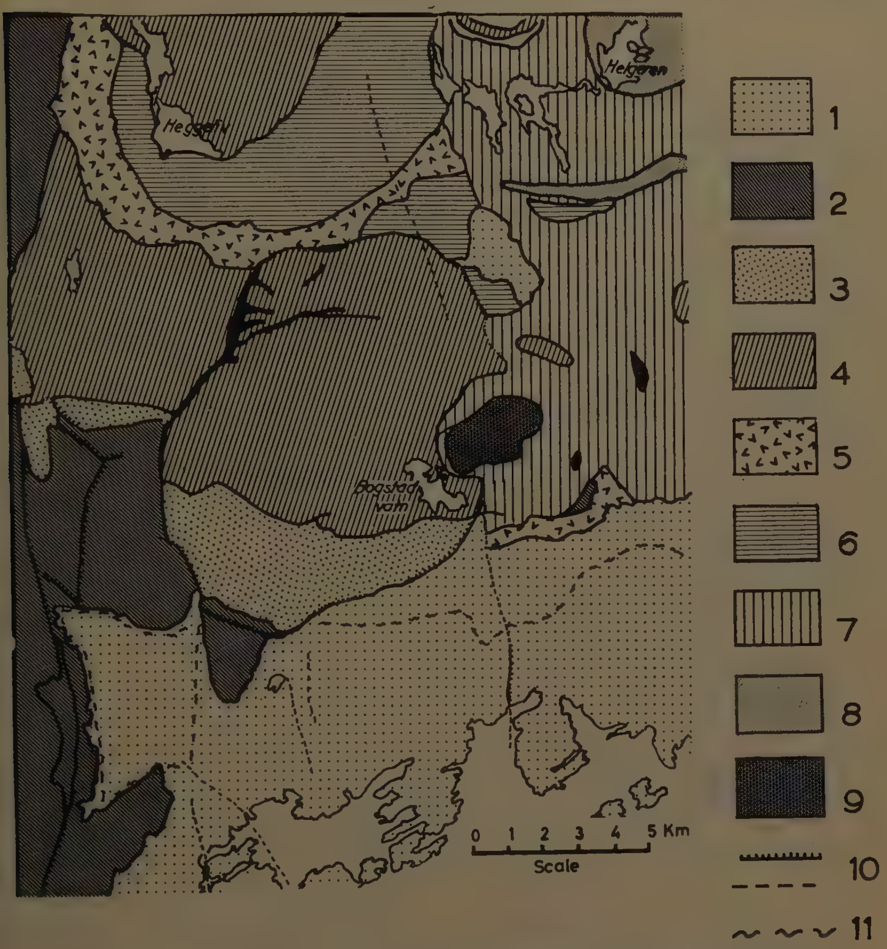


Fig. 1. Geological map of the Bærum cauldron and vicinity (Generalized after Holtedahl and Dons 1952).

Cambrian to lower Devonian sedimentary rocks. Permian supracrustal rocks (with some hypabyssal rocks): 1. Cambrian to lower Devonian sedimentary rocks. 2. Lower rhomb porphyries and B_1+B_2 basalt, 3. B_3-basalt with beds of agglomerate and sandstone, 4. Felsite and syenite porphyry and upper rhomb phorphyries. *Permian intrusive rocks:* 5. Akerite (syenodiorite), 6. Larvikite and Kjelsåsite, 7. Nordmarkite, 8, Ekerite (aegerine granite), 9. Biotite granite. Solid black indicates the large syenitic to granitic (ring) dikes circumscribing the cauldron. 10. Faults and other prominent fractures. 11. Limits of contact metamorphism.

Geology.

The Bærum cauldron is a block, elliptical in plan, that has sunk down inside a ring fault along which a nordmarkitic to granitic dike is emplaced. This Permian cauldron has been described by SÆTHER (1945) and OFTEDAHL (1953, 1960). The details of the geology can be seen on the geologic map (Fig. 1). South of the cauldron lie folded Cambro-Silurian limestones and shales and a Lower Devonian sandstone which have a combined stratigraphic thickness of about 1500 m. The sedimentary rocks are contact metamorphosed adjacent to the cauldron. The Krokskogen lava plateau, composed of Permian basaltic and rhomb porphyry flows, borders the cauldron to the west. Larvikitic plutons lie to the north and nordmarkitic plutons with some granite to the east. The subsided block itself is cut by the eastern nordmarkitic pluton, a fact of particular interest for the gravity interpretation. According to OFTEDAHL (1960), the vertical subsidence, based on the stratigraphic lava sequence, is 1000—1500 m along the southern border. In contrast to the single ring dike elsewhere, the northern end of the subsided block has been intruded by a swarm of nordmarkite dikes as far as 2 km from the marginal fault. The elliptical block has probably subsided into a magma chamber. This subsidence may have been caused by ,,withdrawal of magmatic support'' within the underlying magma chamber or by explosive volcanism whereby the block simultaneously sank into space provided by the extrusion of lava. If the cauldron is underlain by a nordmarkitic to granitic pluton, the mass deficiency of this body should be detectable by means of gravity measurements.

The subsided block is composed of a layered sequence of lava flows and conformable felsite intrusions or ignimbrites which dip northward at 10—30°. The most prominent unit is the B_3 basalt, which attains a maximum thickness of at least 1000 m (SÆTHER, 1945). SÆTHER emphasizes that this is not a plateau basalt but is an extremely heterogeneous unit. It is composed of many small flows with pockets of basaltic clastics and is cut by numerous glassy, basaltic dikes. He calls the B_3 basalt unit a shield volcano. The basalt is visibly metamorphosed. The other rocks in the subsided block are rhomb porphyries and trachytes.

Rock Densities.

The densities of the various rock types were determined by weighing them in air and in water with a balance. Before weighing in water, the samples were allowed to stand in water under a vacuum for 15 minutes. The precision of the method was ± 0.01.

The values obtained appear in Table 1.

Rock Type	No. of Samples	Av. Sp. Gr.
Nordmarkite (Sørkedalen, Vettakollen, Grorud)	6	2.59
Felsites and trachytes in the subsided block ..	5	2.70
Rhomb prophyry (RP_1)	1	2.71
Shale and shaley limestone	4	2.72
Larvikitic rocks (Sørkedalen)	3	2.74
B_3 basalt	3	3.03

The above figures show that the rocks of the subsided block with the exception of the B_3 basalt have about the same density as the rocks outside the cauldron. An average density of 2.73 is used for the country rocks and 2.59 for the nordmarkite plutons.

The low density of nordmarkite is striking; it may be explained by the drusy character of these rocks, first noted by BRÖGGER (1890). The low density of nordmarkite certainly affirms the stoping power of nordmarkite magma. Its lower density would allow it to stope its way into all but rather porous sedimentary rocks.

Gravity Interpretation.

A Bouguer gravity map (Fig. 2) shows the vertical intensity of the earth's gravity field after the Bouguer correction has been made at each station. The Bouguer anomalies represent the interplay of local and regional anomalies, superposed on each other. Local and regional are only relative in denoting any particular areal extent; however, regional anomalies are usually due to larger, deeper features, and local anomalies to shallower ones. The anomalies are caused by the horizontal variation of density distributions associated with geologic features, i.e. folds and faults in layers of different density or plutons whose density is different from that of the country rock. The problem

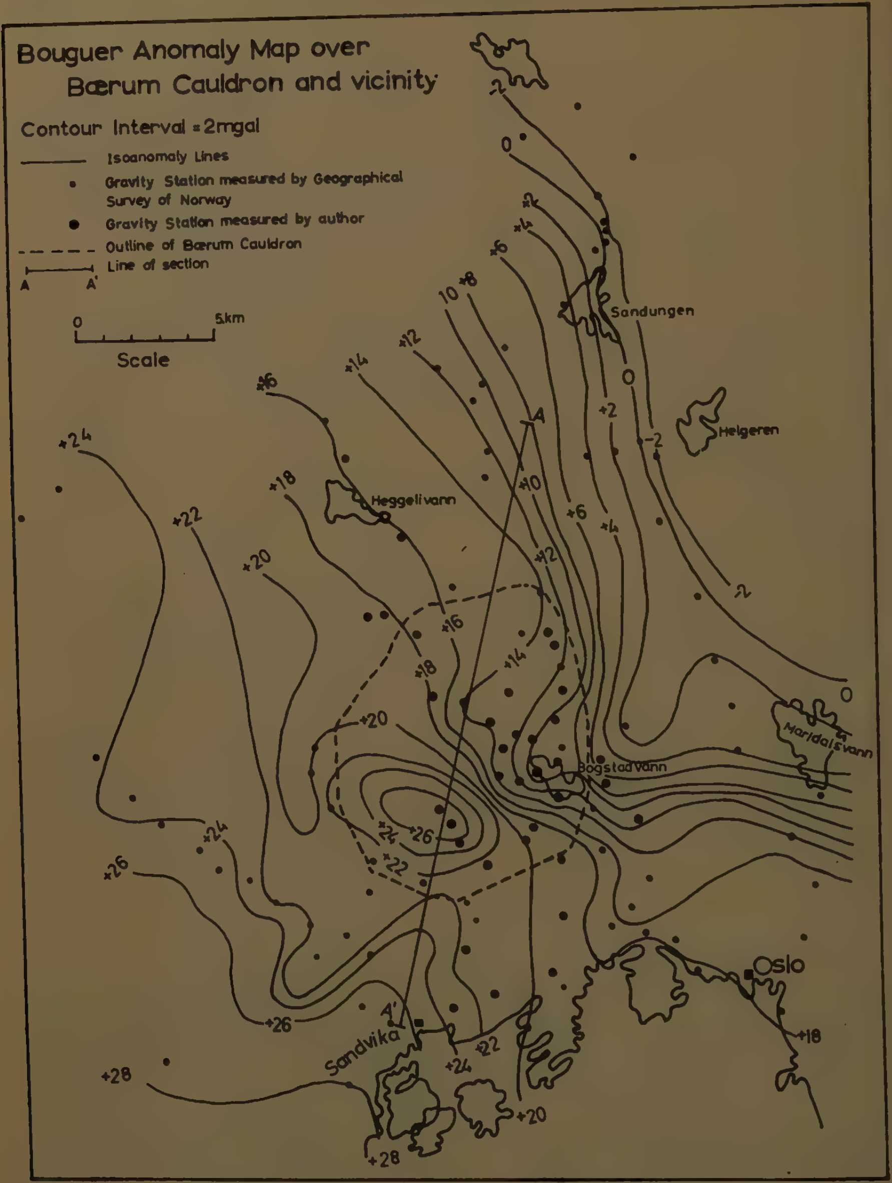


Fig. 2. Bouguer gravity anomaly map.

then, in gravity interpretation, is to isolate the local anomaly and calculate the form for a disturbing mass that corresponds to this anomaly.

A variety of density distributions located at different depths can each cause the same gravity anomaly; therefore, gravity interpretations are inherently ambiguous. The more geological information is known, the more unique the gravity solution becomes.

The geology of the Bærum cauldron indicates that the subsided block is underlain by a nordmarkite pluton. The common association of gravity „lows'' and granitic plutons is well known (ROMBERG and BARNES, 1944; GARLAND, 1950, 1953; BOTT, 1956). The presence of contact metamorphism on the southern rocks of the cauldron (SÆTHER, 1945) suggests that the present depth of the pluton is not great. If the cauldron is underlain by a nordmarkitic to granitic pluton, the mass deficiency of this body should be detectable with gravity measurements. One would expect, therefore, that a „low'' could be detected over the cauldron.

The Bouguer gravity map (Fig. 2) shows no such association. On the contrary, a gravity high is situated over the south end of the cauldron. This „high'' coincides with the outcrop of the B_3 basalt. When the thickness (about 1000 m), high density (3.03), and shallow position of the B_3 basalt are considered, it is not surprising that positive or zero local anomalies occur over the cauldron. Consequently, the effect of the deeper nordmarkite pluton is masked by that of the basalt.

Logically, the next step is to determine the gravity field after the removal of the effect of the basalt. To do this, a N—S profile of the gravity field was drawn through the cauldron (Fig. 3). The regional gravity gradient is expressed in the general slope of the profile outside the limits of the cauldron. The basalt is considered to be a disc dipping northward at 20°, and its attraction is computed at points along the profile. These values are subtracted from the measured gravity at each point to determine a new curve that represents the gravity field if the basalt were replaced by a layer of average density (2.72). The difference between the curve obtained and the regional gradient is the anomaly caused by the pluton at any particular point.

It should be noted that a radius of 4 km was used for the basalt disc. This presupposes that the basalt is intersected in depth by the pluton and does not extend completely across the cauldron. There are

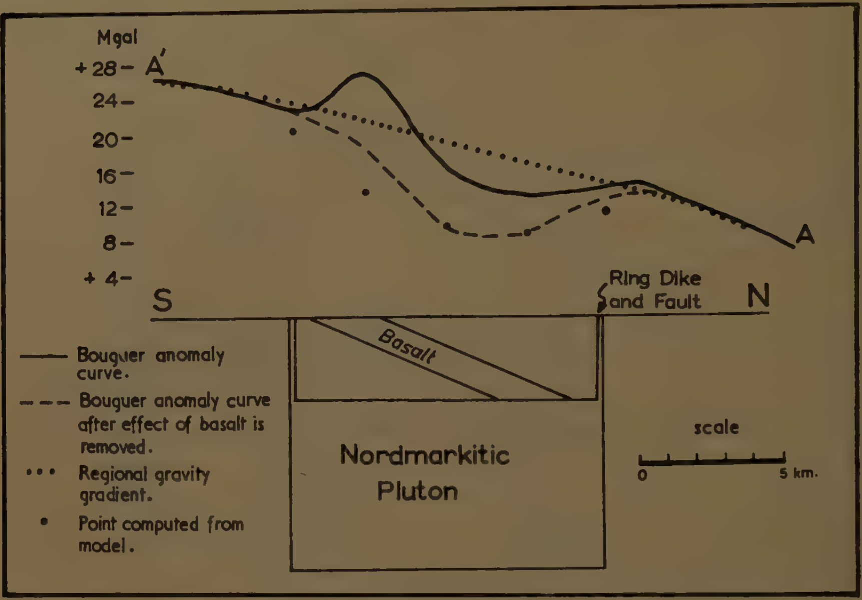


Fig. 3. Profile of gravity field and computed model of nordmarkitic pluton.

two reasons for this supposition: 1) The contact metamorphism suggests a shallow depth of the pluton. 2) The most suitable model can be computed on this basis.

After the effect of the basalt is removed, an anomaly of 10 mgal. appears over the center of the cauldron. This anomaly upholds the geological supposition that a nordmarkite pluton underlies the cauldron. Accordingly, a cylindrical model was used to compute a theoretical anomaly, which was compared with the actual anomaly. A density contrast of 0.14 (2.73 minus 2.59) between surface rocks and nordmarkite was used, and it was assumed that this density contrast exists in depth. A 6-km-deep cylinder of nordmarkite whose top is 3 km below the earth's surface and whose radius is 5 km causes a negative anomaly of 10 mgal.

The attraction of the cylinder at different points along the profile was computed and plotted. Fig. 3 shows that the points near the edge of the cauldron depart from the anomaly curve. This departure is caused by approximate nature of the assumptions used; the data available do not warrant more exact calculations.

A nordmarkitic to granitic pluton that cuts across a small part of the cauldron's east side offers an interesting possibility to check the postulated thickness of the subsided block. This is probably an extension of the pluton underlying the cauldron; therefore, the local anomaly over this extension is a function of the thickness of the subsided block. In this case, the anomaly associated with a 2-km-wide outcrop is the local anomaly, and the gravity field along a 12 km profile furnishes the regional gradient. The maximum negative anomaly over the projection of the pluton east of Bogstadvann is 4 mgal. An average density of 2.82, based on the relative thicknesses of basalt and felsite flows, was used for the subsided block and, again, 2.59 for the intrusion. A half-cylinder approximates the form of the intrusion because it is a semi-rounded projection from a much larger nordmarkitic to granitic pluton that borders the east side of the cauldron. This model, extending to a depth of 3 km, causes a negative anomaly of 4.3 mgal. as compared with measured anomaly of 4 mgal. The close agreement between the two calculations is striking, but they are not completely independent because the average density of the subsided block is necessarily based on depth assumptions. It is, in fact, probably accidental that the two figures agree so well, but, even a somewhat poorer agreement would imply that the depth to the underlying pluton is on the order of 3 km.

Some features of the Bouguer anomaly map require further explanation. The marked positive anomaly over the B_3 basalt around Muren is not expressed to the east where the basalt is also exposed. This fact may be explained by a thinning of the basalt in this direction. For all calculations a constant thickness of 1000 m was used, and departures from this thickness would affect the calculations. A denser net of gravity stations could give information regarding the thickness and would be advisable over the basalt cutcrop and its vicinity.

A traverse northwest from Sandvika in west Bærum includes several relatively low readings that coincide with an area of contact-metamorphosed Cambro-Silurian sedimentary rocks. This „low" is reflected in the southwest trending embayment of the isoanomaly contours 3 km northwest of Sandvika. No plutons are exposed in the area. BOTT (1956) has invoked a large, unexposed granitic intrusion to explain a large negative anomaly over a mineralized region in England. The presence of a granitic intrusion is strongly suggested for this area by the coincidence of an apparent low and the contact metamorphism; however, more gravity stations are needed to define the anomaly.

The most distinctive feature of the gravity map is the „low'' shown by an embayment in the contours over Tryvasshögda. This represents a large „low'' superimposed on the southwest-northeast regional gradient. This „low'' corresponds to the outcrop of a large pluton of nordmarkite and granite and would be another worthwhile project for further study.

Conclusions.

The gravity data corroborate the geological supposition that the Bærum cauldron subsided into a magma chamber. As determined by the geological information used, the cauldron is underlain by a cylindrical pluton whose top is about 3 km below the surface and whose „base'' is 9 km deep. The „base'' is defined as the lower limit of the density contrast between the pluton and country rock; „base'' is not used to indicate whether the pluton or the country rock is terminated at that depth. The calculations indicate a considerable size for the pluton; however, the depth to its top is undoubtedly more accurate than the depth to its „base''.

This regional survey suggests that an unexposed granitic pluton underlies an area in west Bærum northwest of Sandvika and shows a fairly large negative anomaly over the Tryvasshögda nordmarkites. Both areas require more detailed gravity measurements, and a detailed gravity net would also be advisable over the cauldron area. Because of the marked low density of nordmarkite, nordmarkitic areas are particularly suitable for gravity surveys even when in contact with plutons of medium composition, i.e. larvikite. In a geologically complicated area such as the Oslo region, detailed surveys are practically mandatory. These surveys should be planned with a consideration of geological features and an anticipation of the anomalies caused by them.

Acknowledgments.

Research for this project was carried out with the facilities of the Geologisk Museum. The field expenses and rent of the gravimeter were provided by the Geologisk Museum. The project was suggested and arranged by Prof. Tom F. W. Barth. Conservator J. A. Dons discussed the geology and read the manuscript. The Geological Survey of Norway generously made all of their data available and loaned the writer a barometer. Mr. Thorolf Nordby's help in the field was invaluable. Thanks are due to Vækerö—Lövenskiold for permitting the use of their private roads in the area. The writer is deeply grateful to the people and organizations who made this study possible.

REFERENCES.

Bott, M. H. P., 1956: A geophysical study of the granite problem: Geol. Soc. London Quart. Jour. v. 112, p. 45—62. p. 93—117.

— and Masson-Smith, D., 1957: The geological interpretation of a gravity survey of the Alston Block and the Durham Coalfield: Geol. Soc. London Quart. Jour. v. 113, p. 93—117.

Brögger, W. C., 1890: Die Mineralien der Syenitpegmatitgänge der südnorwegischen Augit- und Nephelinsyenite, Zeitschr. für Krystallographie, 16, p. 56.

Garland, G. D., 1950: Interpretations of gravimetric and magnetic anomalies on traverses in the Canadian Shield in northern Ontario, Publications of the-Dominion Observatory, Ottawa, v. 16, no. 1, 57 p.

— 1953: Gravity measurements of the Maritime Provinces, Publications of the Dominion Observatory, Ottawa, v. 16, no. 7. p. 187—243.

Hubbert, M. K., 1948: Gravitational terrain effects of two-dimensional topographic features, Geophysics v. 13, p. 226—254.

Nettleton, L. L., 1940: Geophysical prospecting for oil, McGraw-Hill, New York. 444 p.

Oftedahl, C., 1953: Studies on the igneous rock complex of the Oslo region. XIII. The cauldrons: Vid. Akad. Skr. 108 p.

— 1960: Permian rocks and structures of the Oslo region, *in* Holtedahl, O., Geology of Norway, N.G.U., 208, p. 298—343.

Romberg, F., and Barnes, V. E., 1944: Correlation of gravity observations with the geology of the Smoothing-iron granite mass, Llano County, Texas: Geophysics, v. 9, pl. 79—93.

Sæther, E., 1945: Studies on the igneous rock complex of the Oslo Region. III. The southeastern part of the Bærum—Sörkedal cauldron: Vid. Akad. Skr., No. 59 p.

Manuscript received June 17, 1961.

Printed December 1961.

SYNTHETIC CHLORAPATITES AND EQUILIBRIA IN THE SYSTEM CALCIUM ORTHOPHOSPHATE-CALCIUM CHLORIDE

By

Roger D. Morton, B.Sc., Ph. D., F. G. S.
(Mineralogisk-Geologisk Museum, University of Oslo, Norway.)[1]

Abstract. A study of the system Calcium orthophosphate-calcium chloride in air at atmospheric pressure shows that synthetic chlor-spodiosite melts incongruently at 945° C ± 5° and that a binary eutectic $Ca_2Cl(PO_4)$ — $CaCl_2$, i.e. chlor-spodiosite-hydrophilite, exists at 760° ±5°. Above 1100° C rapid vapourisation of the calcium chloride and the formation of a chloroxyapatite (chlor-voelckerite) prevent the use of the present techniques. The structure of chlor-oxyapatite is distinct from that of chlorine-rich apatite and is apparently very similar to that of certain synthetic mixed-apatites described by Chaudron and Wallaeys (1949). A natural voelckerite with a low chlorine content has been recently found. Artificial chlor-oxyapatite readily forms when pure chlorapatite is heated in air at 1400° C. Experiments performed so far indicate that a loss of chlorine from chlorapatite in air takes place slowly at temperatures as low as 980° C.

Introduction

Problems arising from a study of the genesis of certain phosphate associations in the Precambrian apatite veins of Ödegården, Telemark, S. Norway, revealed the scarcity of information on the relationships between chlor-spodiosite and chlorapatite in the system $Ca_3(PO_4)_2$ – $CaCl_2$.

---

[1] (Present address: Department of Geology, University of Nottingham, England).

Occurrences of relatively pure chlorapatite, $Ca_5Cl(PO_4)_3$, are rare; the chlorapatite localities in S.E. Norway being almost unique. Chlorapatite has also been identified in meteorites; see DANA (1951), vol. II. Spodiosite is a mineral of the composition $Ca_2F(PO_4)$ which was found at the Nyttsta Kran mine in Värmland, Sweden. The chlor-analogue, $Ca_2Cl(PO_4)$, of spodiosite has never been found in the natural state. Calcium chloride, $CaCl_2$, is found naturally as the mineral Hydrophilite; and the low temperature form of calcium orthophosphate or $\beta - Ca_3(PO_4)_2$ as the mineral Whitlockite.

Previous work

Experiments carried out by ST. CLAIRE DEVILLE and CARON (1858) showed that artificial chlorapatite could be synthesised by heating an excess of calcium orthophosphate together with calcium chloride. At the same time it was discovered that on heating a melt containing a much higher proportion of calcium chloride, a compound having the formula $Ca_2Cl(PO_4)$ crystallised from the melt. This compound, on the basis of its analogy to the chemical composition of the mineral Wagnerite, $Mg_2F(PO_4)$, was named Chlor-wagnerite.

Further experiments performed by NACKEN (1912) provided data for the first equilibrium diagram for the binary system $Ca_3(PO_4)_2 - CaCl_2$. Naturally the great advances made during the last 40 years in experimental techniques for mineral synthesis make any such earlier work seem somewhat inaccurate. The precipitated crystals in equilibrium with the melt at any given temperature were extracted manually, using a platinum spoon. The crystals were then examined under the microscope. Obviously such a method would be subject to error owing to the lack of an efficient quenching technique and also to thermal gradients developing within a large volume of melt in a crucible. From this work it was found that the compound $Ca_2Cl(PO_4)$, which Nacken now termed Chlor-calciumwagnerite, melted incongruently at 1040° C into chlorapatite and a liquid. A binary eutectic between $Ca_2Cl(PO_4)$ and $CaCl_2$ was said to exist at 770° C in the mixture $97CaCl_2$ $3Ca(PO_4)_2$ (Wt.%). The melting point of pure chlorapatite was given as 1530° C.

CAMERON and MCCAUGHEY (1911), during experiments on the production of artificial chlorapatite, showed that by dissolving calcium orthophosphate in a calcium chloride melt, a compound was synthesised which they identified as the chlor- analogue of the mineral spodiosite. This product, chlor-spodiosite, is without doubt, the chlor-calciumwagnerite of Nacken.

The most recent study concerning chlor-spodiosite is that of Mackay (1953) on the unit cell and space-group of artificial $Ca_2Cl(PO_4)$. The data for the orthorhombic cell dimensions were said to justify, to some extent, Dana's inclusion of spodiosite in the Wagnerite group.

Experimental methods

General Procedure.

Phase relations within this system were chiefly determined using the quenching method. Mixtures of calcium orthophosphate and calcium chloride in platinum foil envelopes were held in air at constant temperatures until complete equilibrium was obtained, i.e. a state as near to equilibrium as one can hope to obtain in such experiments. The samples were then rapidly quenched to room temperature and the existing phases identified by x-ray and optical methods.

The quenching method was in part supplemented by direct visual observations of melts, using a microscope furnace. This method enabled accurate positioning of the liquidus-solidus curves within calcium chloride rich mixtures whose melts possess a rather low viscosity.

Materials and Mixtures.

Starting materials of the highest available purity were used. The reagents were mixed together in the required proportions and then ground for 30 minutes under purified benzene in an agate mortar. The use of benzene thus prevented any undue contamination by water; which might effect the formation of an OH bearing apatite. The mixtures were then dried at 100° C and tumbled for another 10 minutes.

In the case of mixtures containing more than 50% $CaCl_2$, these were then slowly fused in a platinum crucible. The heating of these mixtures up to the liquidus region was done slowly to prevent loss of charge owing to effervescence of the calcium chloride during melting. The mass was then cooled and reground.

Mixtures containing less than 50% by wt. of calcium chloride were first heated to 850° C in a platinum crucible. The mixture was then chilled and reground. All mixtures were then pressed into small tablets using a cylindrical press. Tablets for heating in tube-furnaces were enclosed in platinum envelopes.

Furnaces and Temperature Control.

Three furnace types were employed during this work. Quenching runs below 1000° C were performed in horizontal "Nicrome"-wound tube-furnaces; whereas a "Super-Kanthal" horizontal tube-furnace was used in runs between 1000° C and 1400° C. The temperatures in the furnaces were controlled with an accuracy of ± 5° C using Chrome-Alumel or Pt/Pt.Rh thermocouples coupled to automatic control units, (type 21c of Joens & Co., Dusseldorf).

During the study of mixtures containing 80 to 99% $CaCl_2$ by wt., a modification of ORDWAY's (1952) microscope furnace was used. The behaviour of melts was studied under the microscope with the melt in the form of a drop suspended in the minute coil of a combined Pt/Pt.Rh furnace-thermocouple. The thermocouple was frequently calibrated with an optical pyrometer (cat. no. 8622-C of Leeds and Northrup Co., Philadelphia). One slight disadvantage of this micro-furnace method must be noted, that is the corrosion of the thermocouple at temperatures above 900° C owing to reaction between the Rhodium component of the alloy and the melt.

Examination of Quenched Samples.

Phase identification in the quenched samples was performed using both optical and x-ray techniques. The x-ray methods involved the use of 9 cm powder-cameras with Mn filtered Fe radiation.

Chlor-spodiosite is easily recognisable for it mostly crystallises as colourless, flattened orthorhombic crystals, with the predominant form being (001); see Plate I, Fig. 1. Some of the chlorspodiosite crystals are acicular in form. These crystals have R.I. $N\beta = 1{\cdot}665$ moderate birefringence and (—) $2V = 70°$ approx. (Mackay gave (—) 2V 75° ± 5°).

The artificial chlorapatite forms long (up to 3 mm) or short hexagonal prisms terminated by bipyramidal faces; see Plate I, Fig. 2.

The apatite varies in colour between sky-blue and greenish-blue and has $N_0 = 1{\cdot}666$ and birefringence around 0·006.

Small amounts of coexisting calcium chloride were difficult to identify owing to its extreme deliquescence. This was in some cases counteracted by sealing the heated sample in a Lindemann glass capillary tube and then using this as the rotating specimen in the powder-camera.

Distinction between the a- and β- forms of calcium orthophosphate was purely on the basis of x-ray techniques. The powder-patterns of chlorapatite and chlor-spodiosite are incorporated in Plate II.

Results

The Region 600° C to 1150° C.

Figure 1, an equilibrium diagram of the system $Ca_3(PO_4)_2$ — $CaCl_2$, summarises the results of the present work on the region between 600° C and 1150° C. The melting point of calcium chloride (772° C) was obtained from "The Handbook of Chemistry and Physics: (1956)": the inversion temperature of calcium orthophosphate (1180° C) from BREDIG, FRANCK and FULDNER (1933).

As is indicated, 13 different compositions were studied and over 70 individual experiments performed, the results being given in Table 1. It has been shown that synthetic chlor-spodiosite melts incongruently at 945° C ± 5° into crystals of chlorapatite and a melt; as distinct from the previously published value of 1040° C. The existence of a binary eutectic between $Ca_2Cl(PO_4)$ and $CaCl_2$ has been confirmed but appears to lie at 760° C ± 5°.

The Region above 1100° C.

The lack of data in Figure 1 in the temperature range above 1100° C is due to the behaviour of mixtures heated in air above this temperature. Firstly a serious factor is introduced by the rapid vapourisation and consequent loss of calcium chloride at these higher temperatures; (the calcium chloride vapourises at lower temperatures, i.e. 850° to 1000° C but at an extremely low rate).

The second factor inhibiting the use of these experimental methods at higher temperatures is perhaps the most interesting one from a geological point of view. This is the previously unrecognised fact that

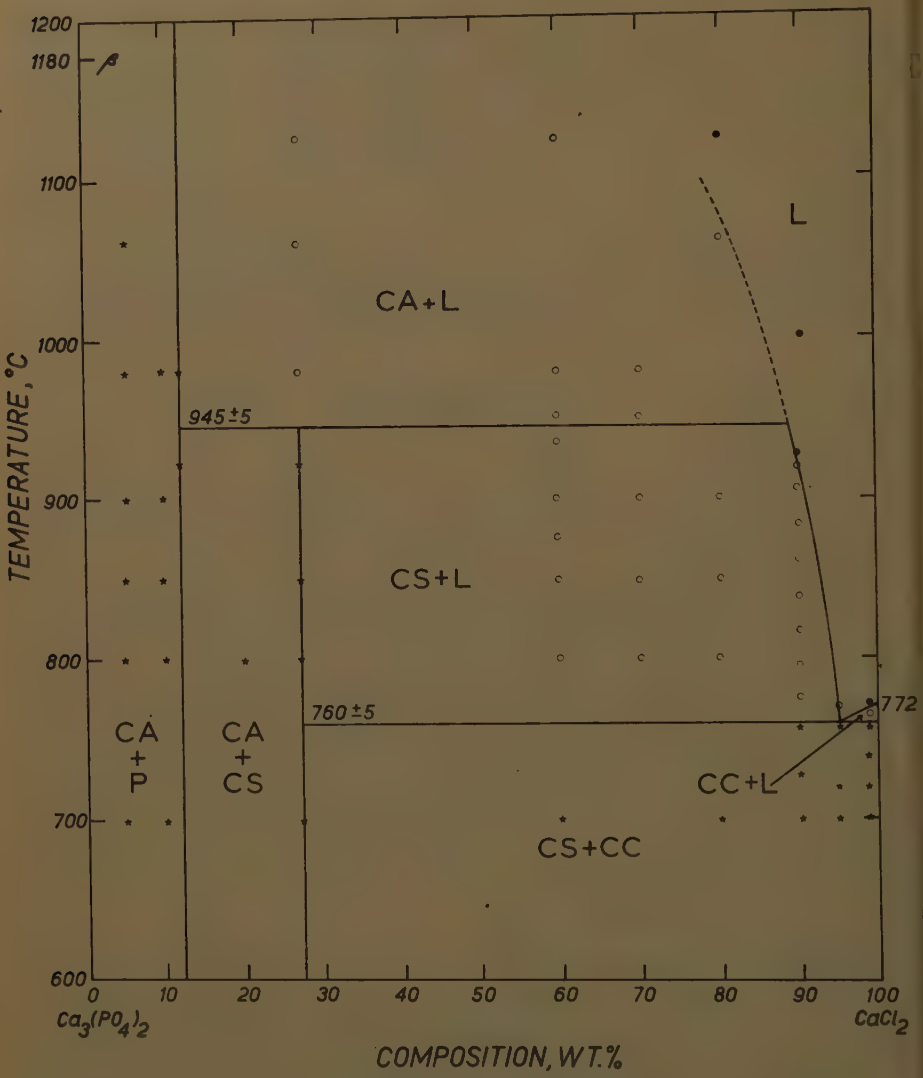


Fig. 1. Equilibrium diagram for the binary system $Ca_3(PO_4)_2$ — $CaCl_2$. Experimental results are indicated by stars (completely crystalline products), circles (crystals in equilibria with liquids) and dots (liquid). CA = Chlorapatite, CS = Chlor-spodiosite, CC = Calcium chloride, P = Calcium orthophosphate, L = Liquid, β = Transition temperature of Calcium orthophosphate.

chlorapatite in air at atmospheric pressure above approximately 1150° C loses most of its chlorine. This loss of chlorine results in the formation of a Chlor-oxyapatite, that is, an oxyapatite (voelckerite) having a low content of chlorine and a formula $Ca_5\,[(O_{\frac{1}{2}} > Cl)\,(PO_4)_3]$. A similar compound has been found to occur naturally (description to be published at a later date) and to the best of the author's knowledge, has not been described previously. A small amount of calcium oxide was found in a few of the samples heated at these higher temperatures. Evidence for this change in chemical composition is based upon analyses of synthetic products. A chlorapatite produced at 980° C from a melt of composition 50% $CaCl_2$ 50% $Ca_3(PO_4)_2$ (Wt.%) was found to contain 5.00% Cl by wt. ; the theoretical content of Cl in a chlorapatite is 6.8%. This apatite was finely ground and the powder heated at 1380° C for 22 hrs.; the resulting product was a well crystallised apatite with only 2.1% by wt. of Cl. The x-ray powder pattern of this heated apatite was in many respects different from that of the original chlorine-rich apatite; see Plate II. The pattern of the heated apatite closely resembles that of synthetic mixed-crystals of chlor-fluor-apatite described by Chaudron and Wallaeys (1949), see also Plate II.

Accompanying these changes in composition and crystal structure is a third and equally marked alteration. The synthetic chlorapatites exhibit a range of colours between colourless, blue and green. However, on heating to temperatures above 1100° C, these colours change to rose-pink, olive-green or pale amber. Such changes must be intimately connected with the alterations in the structural state and the relative proportions of chlorine and oxygen in the molecule.

The appearance of an oxyapatite containing a very low content of chlorine at these higher temperatures clearly indicates that at atmospheric pressure, in the presence of air, the system $Ca_3(PO_4)_2 - CaCl_2$ can no longer be considered as truly binary. Consequently the form of the previously published liquidus-solidus curves are liable to be wrong. At the same time it would appear that published data concerning the melting point of pure chlorapatite in air at atmospheric pressure is in all probability the data for chlor-oxyapatite (chlor-voelckerite): see Nacken (1912) and Trömel (1932). Another point which becomes evident is that any analyses of natural chlorapatite which indicate that water was estimated by loss of weight on heating to

Table 1.

Summary of Experimental Results

(CA = Chlorapatite; CV = Chlor-voelckerite; CS = Chlor-spodiosite; PL = Low temperature form of calcium orthophosphate; PH = High temperature form of calcium orthophosphate; CC = Calcium chloride; Q = Calcium oxide).

Run no.	Duration (hrs.)	Temp. (°C)	Charge (Wt. % $CaCl_2$)	Crystalline Products
10654	168	700	5%	CA + PL
10672	96	800	do	CA + PL
10675	96	855	do	CA + PL
10613	96	900	do	CA + PL
10595	72	980	do	CA + PL
10639	72	1060	do	CA + PL
10644	48	1125	do	CV + PL
10586	48	1200	do	CV + PH
10620	60	1400	do	CV + PH
10655	168	700	10%	CA + PL
10671	96	800	do	CA + PL
10674	96	855	do	CA + PL
10754	48	900	do	CA + PL
10594	72	980	do	CA + PL
10638	72	1060	do	CA + PL
10643	48	1125	do	CV + PL
10592	48	1200	do	CV + PH
10617	60	1400	do	(CV)+PH
10578	48	920	12%	CA
10593	72	980	do	CA
10585	48	1200	do	CV + Q
10627	60	1400	do	PH + (CV)
10803	48	800	20%	CA + CS
10650	168	700	27%	CS
10673	96	800	do	CS
10670	96	855	do	CS
10583	48	920	do	CS
10596	48	980	do	CA
10635	72	1060	do	CA
10642	48	1125	do	CA
10589	48	1200	do	CV
10698	48	1320	do	CV
10624	60	1400	do	PH
10659	168	700	60%	CS + CC
10696	96	800	do	CS + CC

Table 1, continued.

Rnm no.	Duration (hrs.)	Temp. (°C)	Charge (Wt. % $CaCl_2$)	Crystalline Products
10690	96	855	do	CS + CC
10777	48	875	do	CS + CC
10817	36	885	do	CS + CC
10674	48	900	do	CS + CC
10859	48	935	do	CS + CC
10611	84	980	do	CA + CC
10647	48	1125	do	CA + (Q)
10649	72	1060	do	CA + (Q)
10699	48	1320	do	CV
10623	60	1400	do	PH
10693	96	800	70%	CS + CC
10697	96	855	do	CS + CC
10575	48	900	do	CS + CC
10804	48	950	do	CA + CC
10616	84	980	do	CA + CC
10658	168	700	80%	CS + CC
10692	96	800	do	CS + CC
10700	96	855	do	CS + CC
10584	40	900	do	CS + CC
10648	72	1060	do	CA + (Q?)
10640	48	1125	do	CA + (Q?)
10608	60	1400	do	PH
10580	40	900	90%	CS + CC

about 1200° C will possibly be unreliable. This would be especially applicable to those performed prior to the introduction of the Penfield method for water determination in 1894.

The exact relationships between chlorapatite and chlorvoelckerite are being studied in greater detail and will be the subject of a later publication. It is however worthy of note that the experimental results so far obtained indicate that the loss of chlorine from a chlorapatite heated in air can take place slowly at temperatures as low as 980° C.

Acknowledgements

The author is indebted to Professor T. F. W. Barth and Dr. H. Neumann for their constant assistance and encouragement during the course of this investigation. Thanks are expressed to Ing. B. Bruun for the analytical work included in this paper, to Cand.real. O. H. J. Christie for his advice concerning the micro-furnace and other aspects of this work, and to Dr. A. L. Mackay who kindly supplied chlor-spodiosite for reference purposes.

LITTERATURE CITED

BREDIG, M. A., FRANCK H. H. and FULDNER H., 1932. Beiträge zur Kenntnis der Kalk-Phosphorsaure-Verbindungen I. Zeit. Elektrochem., vol. 38, P. 158.

CAMERON F. K. and McCAUGHEY W. J., 1911. Apatite and Spodiosite. Jour. Phys. Chem., vol. 15, p. 463.

CARON H. and St. CLAIR-DEVILLE M., 1858. Mémoire sur l'apatite, la wagnerite et quelques espèces artificielles de phosphates métalliques. Compt. Ren., vol. 47, p. 985.

CHAUDRON G. and WALLAEYS. R., 1949. Synthèse des apatites par réaction dans l'état solide. Bull. Soc. Chim. France, p. D-132.

MACKAY, A. L., 1953. The unit cell and space-group of chlor-spodiosite (Ca_2PO_4 Cl). Min. Mag., vol. XXX, no. 222, p. 166.

NACKEN R., 1912. Ueber die Bildung des Apatits. I. Centralblatt für Mineralogie logie, p. 545.

ORDWAY F., 1952. Techniques for growing and mounting small single crystals of refractory compounds. Paper no. 2299, Jour. of Research of the Nat. Bur. of Standards, vol. 48, no. 1, p. 152.

PALACHE C., BERMAN H. and FRONDEL C., 1951. The system of mineralogy of James Dwight Dana and Edward Salisbury Dana. 7th edition, vol. II.

TRÖMEL G., 1932. Untersuchungen über die Bildung eines halogenfreien Apatits aus basischen Calciumphosphaten. Zeits. Physikal. Chem., Abt. A, vol. 158, p. 422.

Manuscript received July 12, 1960.

Printed December 1961.

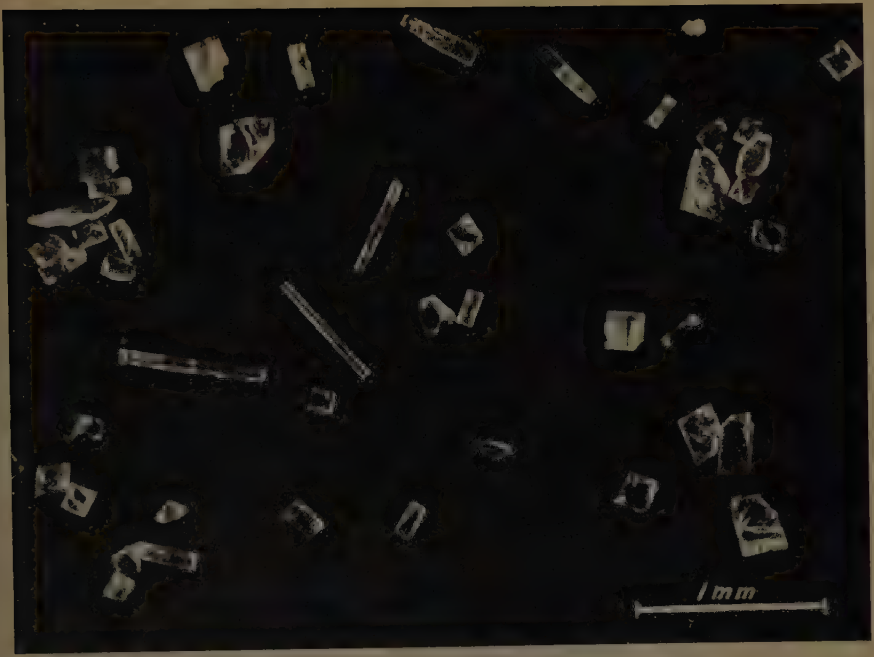


Fig. 1. Synthetic chlor-spodiosite crystals exhibiting both the characteristic flattened (001) form and the rarer acicular habit.

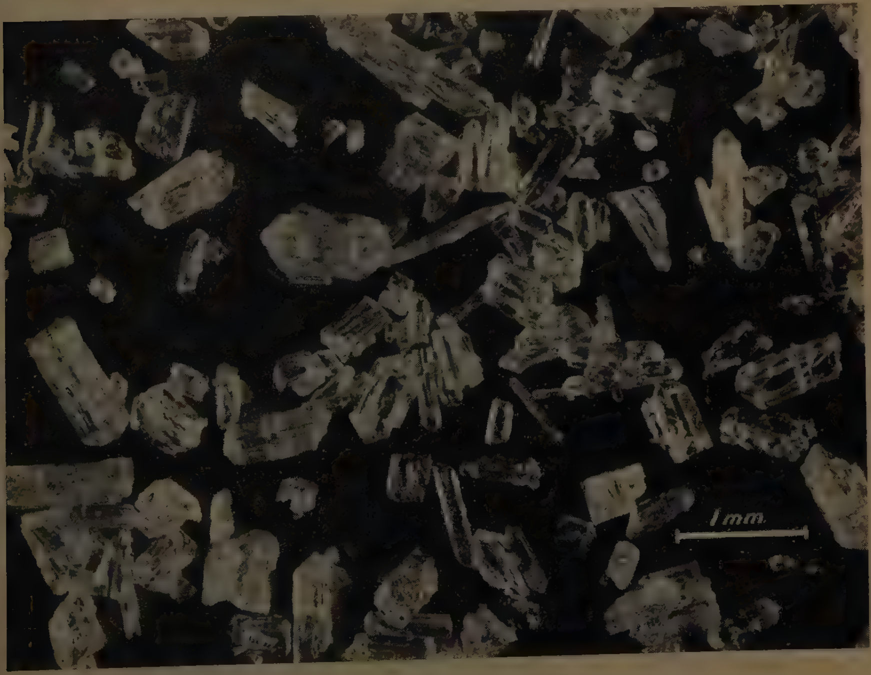
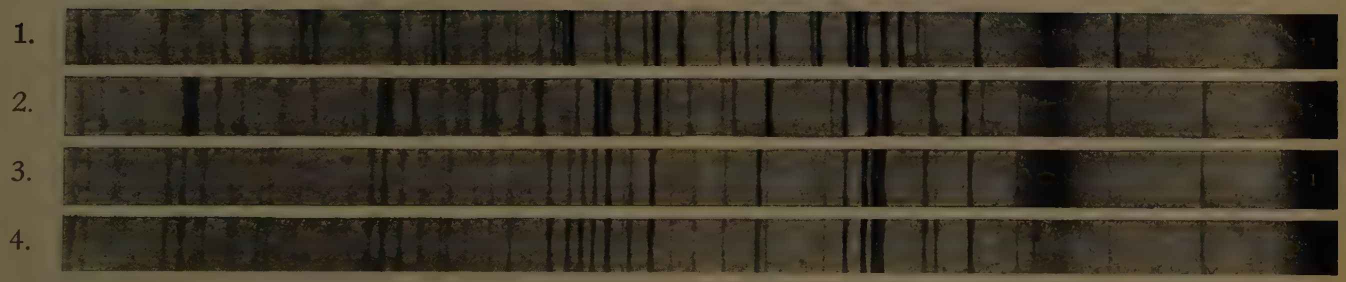


Fig. 2. Synthetic chlorapatite showing the typical development of short hexagonal prismatic crystals terminated by bipyramidal faces.

PLATE 2



X-ray powder diffraction photographs of synthetic apatites and chlor-spodiosite taken on a Nonius camera of diameter 11,459 cm using CuK radiation. Exposure time 3 hrs. 1. Synthetic chlor-spodiosite; 2. Synthetic chlorapatite with 5% by wt. Cl: 3. Synthetic chlor-voelckerite containing 2,1% by wt. Cl produced by heating chlorapatite (pattern no. 2) to 1380° C. for 22 hrs: 4. Synthetic mixed chlor-fluor apatite with F-Cl approximately 60% F.

CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 9. On the occurrence of two rare phosphates in the Ødegården Apatite mines, Bamble, South Norway

1. A variety of woodhouseite
2. Whitlockite

By

Roger D. Morton, B.Sc., Ph. D., F.G.S.
(Mineralogisk-Geologisk Museum,
University of Oslo, Norway.[1])

Abstract. A description is given of varieties of woodhouseite and whitlockite found in specimens from the apatite-rich veins cutting the Precambrian rocks at Ødegårdens verk, Bamble, S. Norway. The minerals occur, together with quartz, in pockets within a matrix of chlor-hydroxy-oxyapatite. Physical and optical data are given for both minerals, together with x-ray powder diffraction data. A semi-quantitative spectrographic analysis of the whitlockite is included.

The woodhouseite is a purple variety with lattice dimensions of a_h 7,001, c_h 16,265, c/a 2,325, a_r 6,76 and $\angle\, a$ 62° 22'; its optical constants are $N_e = 1{,}669$ (±0,003), $N_o = 1{,}662$ (± 0,003), $N_e - N_o$ 0,007 (± 0,003), uniaxial positive (or occasionally biaxial with 2V of 10° to 20°). An unidentified alteration product after woodhouseite is also described.

The whitlockite is a green variety, containing approximately 0,3% V_2O_5, with optical constants: $N_o = 1{,}620$ (± 0.003), $N_e = 1{,}623$ (± 0,003), $N_e - N_o$ 0,003 (± 0,003), uniaxial negative.

It is concluded that the minerals in these pockets represent either final residual products of crystallisation within the apatite vein or that they were formed during a later period of metamorphism, metasomatism and hydrothermal activity which may have affected the apatite veins.

[1] Present address: Department of Geology, University of Nottingham, Nottingham, England.

General Introduction

At Ødegårdens verk in Bamble, Telemark, South Norway, phosphate-bearing veins, up to 4 m thick, cut the Precambrian metamorphic rocks of the Kongsberg—Bamble formation. The metamorphic suite in the area comprises garnetiferous amphibolites, quartzites, scapolite-hornblende metagabbro (the 'Ødegårdite' of W. C. Brøgger) and scapolite-amphibolites. The most common type of phosphate-bearing vein has central lensoid bodies composed of phosphate minerals and rutile. These lenses are surrounded by a dense mass of coarse-grained phlogopite, enstatite (often pseudomorphed by talc), scapolite and phosphates. From preliminary investigations it has been concluded that a later group of granitic pegmatites cuts these phosphate-rich veins.

A list of some of the minerals found in these veins has already been published (see NEUMANN, JØSANG and MORTON 1960 p. 15). The predominant phosphates occurring within these bodies are apatites, (generally chlor-varieties). The area was extensively mined for the apatite between 1872 and 1918.

During a study of the minerals from these veins, the author had access to W. C. Brøgger's original collection from the old apatite mines. Two small specimens labelled "Apatite med grøn Varietet, Ødegården, Bamble", (i.e. Apatite with green variety) were examined in detail and found to contain as well as apatite, whitlockite and occasional tiny crystals of a variety of woodhouseite.

The samples containing these two rare phosphates were obviously collected from one of the apatite-rich veins during the mining operations. In both samples the host-matrix is a coarse grained massive chlor-hydroxy-oxyapatite (i.e. chlor-hydroxy-voelckerite). Small pockets within the apatite are filled by quartz, whitlockite and woodhouseite; see Plate I, fig. 1. The quartz generally occurs around the periphery of the pocket as colourless, transparent, euhedral prismatic crystals up to 5 mm in length and 1 mm broad. Most of the space in these pockets is occupied by whitlockite. The whitlockite occurs as coarsely crystalline masses. Always within the whitlockite, and never within the apatite, are tiny groups of woodhouseite crystals. In the specimens occasional chalcopyrite grains are seen in later veins which cut the apatite.

1. Woodhouseite from Ødegården

INTRODUCTION.

Woodhouseite, a very rare member of the Beudantite group (Alunite-Woodhouseite group of STRUNZ 1957) has only once before been found. This mineral, having the composition — [$CaAl_3(PO_4)(SO_4)(OH)_6$], was described by LEMMON (1937) from White Mountain, Mono County, California. There it is found together with masses of lazulite in vugs within quartz veins which cut andalusite deposits. Topaz, augelite, tourmaline, barite, muscovite and pyrophyllite are also found in these veins.

The type material from White Mountain is described as colourless, white, or flesh-coloured, striated rhombohedral crystals which are transparent or translucent. Their habit is predominantly pseudo-cubic. The optical character is uniaxial positive, but certain crystals were composed of six radial biaxial segments having an axial angle $2V \doteq 20°$ maximum. The birefringence is 0,011 and R.I. $n\alpha = 1{,}636$, $n\beta = 1{,}638$ and $n\gamma = 1{,}647$ ($\pm$ 0,003). An excellent basal cleavage parallel to (0001) was observed. The original mineral was found to be hexagonal with cell dimensions of $a_h = 6{,}961$; $c_h = 16{,}27$; $c/a = 2{,}338$; $a_{rh} = 6{,}75$; $\alpha = 62° 04'$ (PABST 1947).

DESCRIPTION

a. *Physical and optical properties.*

The woodhouseite from Ødegården occurs as a few tiny clusters of euhedral crystals whose maximum dimensions are 1,5 mm. These exhibit a pseudo-cubic form $(10\bar{1}2)$. The crystal faces are highly striated in all the specimens. Their colour varies from a pale lilac to a deep purple. In most crystals there is a rough zoning, with the lighter lilac-coloured woodhouseite in the central parts, grading outwards into the dark purple around the outer periphery of the crystals. All the crystals are completely transparent, possess a very poor cleavage and generally have an irregular fracture. Their lustre is vitreous and H = ca. 4.

Rarely the woodhouseite is altered to a cream-coloured or white, very fine-grained, powdery substance; (see below).

The optical properties, obtained from the powdered mineral, are as follows. Very thin fragments are colourless but larger grains i.e. $> 0{,}25$ mm exhibit a very weak pleochroism in shades of pale lilac. It is uniaxial positive in most grains but some occasionally possess a positive biaxial character with an axial angle $2V = 10° \text{ to } 20°$. The indices of refraction, measured at 20° C using monochromatic Na light are: $N_e = 1{,}669$ (0,003); $N_o = 1{,}662$ (0,003); $N_e - N_o = 0{,}007$ (0,003). The biaxial fragments show low-order dark interference colours between crossed nicols.

b. *Crystallographic observations from powder diffraction patterns.*

Table 1 serves as a comparison between the observed spacings and intensities (down to $d = 1{,}293$) for woodhouseite from Ødegården and those of woodhouseite from the White Mountains. The observed intensities of lines in the new woodhouseite pattern were obtained by comparison of peak-heights on a photometer-curve. The somewhat different observations of various authors concerning the data for material from White Mountain is also included in Table 1; (see YGBERG 1945, McCONNELL 1942 and PABST 1947).

Table 2 serves as a supplement to the preceding table and compares the lines of $d < 1{,}293$ observed in the pattern of the mineral from Ødegården with those previously published.

The corresponding lattice dimensions were calculated from the data given in Tables 1 and 2. The powder pattern was obtained with a 9 cm Debye-Schaerer camera using Fe radiation with an Mn filter. Table 3 shows that the lattice dimensions of the woodhouseite from Ødegården are in relatively close agreement with the results of Lemmon, Pabst and Ygberg. It is however noticeable that all these dimensions, with the exception of the hexagonal 'c' axis, differ slightly from the data obtained from the earlier studied materials.

c. *Conclusions concerning the woodhouseite.*

From the observations given above, it is concluded that the mineral found at Ødegården is definitely a member of the Beudantite group possessing properties suggest it to be a variety of woodhouseite. However, owing to the lack of material for chemical analysis, no definite conclusions can be reached concerning the causes of small, but distinct, optical and structural variations from previously recorded data on woodhouseite.

Table 1. Observed spacings and intensities in powder diffraction patterns of woodhouseite for intervals higher than d_{hkl} 1.29.

Locality: White Mtn., N. Inyo Range, Cal. USA								Ødegården		
Author: Pabst (1947)				Ygberg (1945)		McConnell (1942)		Morton		
Radiation: Cu				Cu		Fe		Fe		
hk$\bar{i}$l	$h_rk_rl_r$	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	hkil
								0.5	6.229	
10$\bar{1}$1	100	.30	5.66–	m	5.73	<0.8	5.61	0.7	5.688	10$\bar{1}$1
0003	111									0003
01$\bar{1}$2	110	.40	4.84+	m	4.86	0.8	4.81	0.6	4.834	01$\bar{1}$2
								0.2	4.139	
								0.2	3.844	
						<0.8	3.83	3.9	3.704	
11$\bar{2}$0	10$\bar{1}$	.90	*3.48+	s	3.51	1.7	3.46			11$\bar{2}$0
10$\bar{1}$4	211									10$\bar{1}$4
		.20	3.27^1			3.3	3.21^1	0.8	3.241^1	
02$\bar{2}$1	11$\bar{1}$					3.3	2.96			02$\bar{2}$1
11$\bar{2}$3	210	2.45	*2.93–	s+	2.99	10.0	2.91	10.0	*2.939	11$\bar{2}$3
01$\bar{1}$5	221	.15	2.84					0.7	2.832	01$\bar{1}$5
20$\bar{2}$2	200									20$\bar{2}$2
0006	222	.45	2.70–	w	2.72	1.7	2.68	1.7	2.721	0006
		.15	2.56 §							
02$\bar{2}$4	220	.40	*2.42	w	2.40	1.7	2.42	1.6	2.436	02$\bar{2}$4
						3.3	2.37			
21$\bar{3}$1	20$\bar{1}$									21$\bar{3}$1
20$\bar{2}$5	311							1.7	2.216	20$\bar{2}$5
12$\bar{3}$2	21$\bar{1}$ }	2.10	{2.19+	s+	2.18	5.0	2.198	3.8	2.145	12$\bar{3}$2
10$\bar{1}$7	322 }		{2.16–			6.7	2.15	6.1	2.134	10$\bar{1}$7
11$\bar{2}$6	321									11$\bar{2}$6
		.20	2.09+^{1}			3.3	2.068^1	0.5	2.088^1	
30$\bar{3}$0	2$\bar{1}\bar{1}$									30$\bar{3}$0

Table 1, continued.

Locality: White Mtn., N. Inyo Range, Cal., USA								Ødegården		
Author: Pabst (1947)				Ygberg (1945)		McConnell (1942)		Morton		
Radiation:		Cu		Cu		Fe		Fe		
hkil	$h_rk_rl_r$	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	hkil
$21\bar{3}4$	310	.20	1.99 —	w–	2.01					$21\bar{3}4$
$01\bar{1}8$	332							0.5	1.932	$01\bar{1}8$
$03\bar{3}3$	$22\bar{1}$	2.30	*1.885	s+	1.89	3.3	1.92	6.8	*1.894	$03\bar{3}3$
$30\bar{3}3$	300					8.5	1.886			$30\bar{3}3$
$12\bar{3}5$	320									$12\bar{3}5$
$02\bar{2}7$	331	.10	*1.844							$02\bar{2}7$
0009	333	.20	*1.804							0009
$22\bar{4}0$	$20\bar{2}$	1.90	*1.741	s	1.73	6.7	1.74	5.7	*1.749	$22\bar{4}0$
$20\bar{2}8$	422	.20	*1.685	w–	1.68					$20\bar{2}8$
$13\bar{4}1$	$21\bar{2}$									$13\bar{4}1$
$22\bar{4}3$	$31\bar{1}$									$22\bar{4}3$
$31\bar{4}2$	$30\bar{1}$	.40	* 1.638			1.7	1.638			$31\bar{4}2$
$21\bar{3}7$	421	?	B 1.621	w–	1.62					$21\bar{3}7$
						1.7	1.619			
$03\bar{3}6$	330									$03\bar{3}6$
$30\bar{3}6$	$41\bar{1}$									$30\bar{3}6$
$11\bar{2}9$	432	.40	*1.602	w–	1.60	1.7	1.60	1.5	1.609	$11\bar{2}9$
$10\bar{1}.10$	433	.10	1.568			1.7	1.573			$10\bar{1}.10$
$13\bar{4}4$	$32\bar{1}$	.10	1.546							$13\bar{4}4$
$12\bar{3}8$	431	.45	*1.519	w–	1.51	1.7	1.515			$12\bar{3}8$
$40\bar{4}1$	$3\bar{1}\bar{1}$									$40\bar{4}1$
$31\bar{4}5$	410	.60	1.487	w	1.48	3.3	1.486	1.6	1.489	$31\bar{4}5$
$04\bar{4}2$	$2\bar{2}2$							2.6	1.479	$04\bar{4}2$
$22\bar{4}6$	420	.90	*1.466	w	1.46	3.3	1.464			$22\bar{4}6$
$01\bar{1}.11$	443									$01\bar{1}.11$

Table 1, continued.

Locality: White Mtn., N. Inyo Range, Cal. USA								Ødegården		
Author: Pabst (1947)				Ygberg (1943)		Mc Connell (1942)		Morton		
Radiation: Cu				Cu		Fe		Fe		
hkil	$h_rk_rl_r$	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	hkil
$02\bar{2}.10$	442	1.55*	1.432	m	1.43	5.0	1.426	3.7	1.437	$02\bar{2}.10$
$40\bar{4}4$	400	.25*	1.416					1.0	1.418	$40\bar{4}4$
$32\bar{5}1$	$30\bar{2}$									$32\bar{5}1$
$04\bar{4}5$	$33\bar{1}$	.55B	1.366	w—	1.36			1.2	1.364	$04\bar{4}5$
$23\bar{5}2$	$31\bar{2}$									$23\bar{5}2$
$13\bar{4}7$	430									$13\bar{4}7$
000.12	444									000.12
$03\bar{3}9$	441	.40*	1.345	w—	1.34			1.2	1.349	$03\bar{3}9$
$30\bar{3}9$	522									$30\bar{3}9$
$20\bar{2}.11$	533	.40B	1.332	w	1.31			0.9	1.324	$20\bar{2}.11$
$21\bar{3}.10$	532									$21\bar{3}.10$
$41\bar{5}0$	$3\bar{1}\bar{2}$									$41\bar{5}0$

* Lines used in cell dimension calculations.
§ Unidentified line; contamination?
[1] β lines.

Table 2. Observed lines in woodhouseite powder diffraction patterns for the interval d_{hkl} 1.29 to d_{hkl} 0.90.

N. Inyo Range, California				Ødegården	
Pabst (1947)		Ygberg (1945)		Morton	
Cu rad.		Cu rad.		Fe rad.	
I	d_{hkl}	I	d_{hkl}	I	d_{hkl}
4	1,282	m	1,27	2,7	1,286
1	1,262				
2	1,225				
5	1,192	m	1,19	2,0	1,194
				1,2	1,171
4	1,164	w+	1,16	1,5	1,166
3−	1,138	w	1,13	0,5	1,141
2	1,124				
2	1,106			1,0	1,110
4	1,098	w−	1,10	1,0	1,103
4	1,078	w+	1,07		
2+	1,067				
2+	1,056				
2	1,036				
2	1,031				
3	1,007			0,9	1,010
3	1,003	w	1,00		
2−	0,962				
5	0,953	w+	0,95		
4	0,945	w	0,94		
5	0,935	w+	0,93		
2−	0,930				
3+	0,913	w−	0,91		
5	0,903	w	0,90		

Table 3. Lattice dimensions of woodhouseite.

Author	Hexagonal axes			Rhombohedral axes	
	a_h	c_h	c/a	a_r	∠ *a*
Lemmon (1937)	—	—	1.170 (0,5 × 2,340)	—	—
Ygberg (1946)	6,93	16,3	2,352	6,75	*61°75'
Pabst (1947)	6,961	16,27	2,338	6,75	62°04'
Morton	7,001	16,265	2,325	6,76	62°20'

* Value said to be 61°48' by Pabst (1947).

d. *The alteration product after woodhouseite.*

A feature rarely seen in these specimens is the alteration of the woodhouseite crystals to a white or cream-coloured powdery aggregate. This alteration is either confined to the periphery of the crystal or to its core. The scarcity of material made it impossible to identify this alteration product. However, certain observations were made: —

The mineral is in the form of an extremely fine-grained, powdery mass which under the microscope is seen to be composed of completely anhedral crystals. The refractive index lies in the region of 1,7 and the birefringence is moderate. An x-ray powder diffraction run gave a very weak film in which the three strongest lines lay at d_{hkl} values of 4,16 (strongest), 3,75 (2nd strongest) and 3,41 (3rd strongest).

2. Whitlockite from Ødegården

INTRODUCTION.

Whitlockite, the natural form of $\beta - Ca_3(PO_4)_2$ i.e. the low-temperature polymorph of tricalcium phosphate, was originally found as a hydrothermal mineral associated with siderite, quartz, apatite, ludlamite, fairfieldite, xanthoxenite and triphyllite in a pegmatite at Palermo mine near N. Groton, New Hampshire, U.S.A. (FRONDEL 1943). It has also been found in phosphate-rock deposits on the island of Sombrero near the Virgin Islands (JULIEN 1865 and FRONDEL 1943), on the Los Monges keys in the gulf of Maracaibo (FRONDEL 1943) and on Table Mountain, Curaçao in the West Indies (KLOOS 1888 and FRONDEL 1943). Whitlockite was also recorded from cave-deposits at Sebdou, Oran, in Algeria (Bannister 1947) and from certain slags (TRÖMEL 1932 [DANA]; TRÖMEL and MÖLLER 1932; BREDIG, FRANCK and FÜLDNER 1932; ibid 1933). As far as the author is aware, there is no previous record of whitlockite in Norway.

The whitlockites previously described were transparent to translucent, colourless, white, grey or yellowish. The lustre was said to be vitreous or subresinous. In transmitted light they were colourless with N_o = 1,607 to 1,629, N_e = 1,604 to 1,626, birefringence 0,003 and uniaxial negative; (DANA 1951).

DESCRIPTION.

a. *Physical and optical properties.*

The whitlockite from Ødegårdens verk was found as coarse- and fine-grained, crystalline massive aggregates occasionally showing fragmentary, striated crystal-faces. The mineral is highly variable in colour and ranges from colourless, in a few cases, to a fine pale emerald green in most parts. In all the specimens it is completely transparent. There is no distinct zonation of the colour and it appears to be highly irregular in its distribution. The mineral exhibits no cleavage and has a subconchoidal fracture. Its lustre is vitreous and the hardness approximately 5.

The optical properties, measured using the powdered mineral, are as follows. It is completely transparent and colourless in transmitted light, uniaxial negative with $N_o = 1{,}620$ ($\pm$ 0,003), $N_e = 1{,}623$ ($\pm$ 0,003) and birefringence 0,003.

b. *Observations from x-ray powder diffraction diagrams.*

Table 4 compares data for observed spacings and intensities in the whitlockite from Ødegården and the data for previously described whitlockites. The Θ values published by BREDIG et al (1932) have been recalculated to d_{hkl} values for the purpose of direct comparison.

It will be seen from the above mentioned table that the data for the mineral from Ødegården agrees well with that from the earlier whitlockites.

c. *Spectrographic analysis.*

Owing to the scarcity of material, no detailed chemical analysis was performed. However, a semiquantitative spectrographic analysis of the bright green whitlockite was kindly done by Prof. Ivar Oftedal. The results were as follows:

Main components — P, together with Ca and less Mg.

Traces — Si, Fe, Mn, Al, Na and V. (The vanadium line was remarkably strong and appears to indicate a concentration of around 0,3% V_2O_5).

It is noteworthy that this is a vanadium-rich whitlockite and it is possible that the unusual green colour may be caused by this vanadium content.

Table 4. Observed spacings and intensities in powder diffraction patterns of natural and synthetic whitlockite.

Sebdou, Alg.		N. Groton, NH.		Palermo		Artificial prod.		Ødegården	
A.S.T.M. Ind.		Frondel (1943)		Fisher 1958		Bredig et al 1932		Morton	
Radiation:		Cu		Fe		Cu		Fe	
I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}
		2.0	8.03			m	8.815		
		5.0	6.55	Only the		ms	6.505	0.4	6.323
60	5.2	8.0	5.24	three stron-		s	5.242	2.3	5.103
				gest lines		w	4.392		
				published.					
		4.0	4.07			ms	4.111	1.5	4.003
50	3.43	7.0	3.45			vw	3.831	3.8	3.401
		1.0	3.35			s	3.468	1.0	3.309
70	3.17	9.0	3.21	7.0	3.16	vs	3.221	6.3	3.168
		1.0	3.02			w	3.025	1.3	2.961
100	2.85	10.0	2.88	10.0	2.85	vs	2.897	10.0	2.843
		3.0	2.75			m	2.838	2.4	2.719
		1.0	2.68			m	2.725	0.6	2.680
70	2.59	9.0	2.60	8.0	2.58	vs	2.605	8.7	2.576
		2.0	2.52			mw	2.533	1.0	2.525
20	2.39	2.0	2.41			mw	2.421	1.5	2.489
								1.0	2.379
						m	2.270	0.6	2.340
20	2.24	2.0	2.25			m	2.208	1.2	2.235
		2.0	2.19			m	2.169	1.6	2.168
20	2.15	2.0	2.16					2.2	2.139
		1.0	2.07			m	2.088		
20	2.05	2.0	2.04			m	2.043		
20	2.01	1.0	2.00			m	2.009	2.3	2.011
						s	1.951	1.1	1.973
60	1.92	6.0	1.93			s	1.924	5.1	1.913
60	1.86	5.0	1.88					2.2	1.873
						m	1.850	2.9	1.857
		1.0	1.82			m	1.819	1.3	1.822
20D	1.79	3.0	1.77			m	1.789	2.2	1.755
100	1.71	7.0	1.72			vs	1.736	5.5	1.709
		1.0	1.70			w	1.694	1.4	1.694
		2.0	1.67			mw	1.645	1.7	1.666
		1.0	1.63					0.8	1.623
		1.0	1.60			m	1.612	0.8	1.609
								0.6	1.588

Table 4, continued.

Sebdou, Alg.		N. Groton, NH.		Palermo		Artificial prod.		Ødegården	
A.S.T.M. Index		Frondel (1943)		Fisher (1958)		Bredig et al (1932)		Morton	
Radiation:		Cu		Fe		Cu		Fe	
I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}	I	d_{hkl}
60	1.55	5.0	1.55			s	1.562	3.1	1.537
						vw	1.533		
						w	1.513	0.9	1.505
						mw	1.478	0.4	1.489
		1.0	1.46			mw	1.458	0.8	1.453
		1.0	1.43			mw	1.440	1.3	1.427
		1.0	1.40			m	1.418	1.8	1.398
		1.0	1.38			m	1.392	0.5	1.376
								0.9	1.292
						etc.		0.5	1.291
								0.6	1.278
20	1.26	2.0	1.25			(Data calc.		1.3	1.249
50	1.24	2.0	1.24			from published		0.5	1.241
						values of Θ.)		0.8	1.232
								0.6	1.210
50	1.18	1.0	1.18					0.7	1.178
								0.4	1.172
		1.0	1.15					0.5	1.152
								0.5	1.141
								0.4	1.121
60	1.12	3.0	1.11					1.0	1.115
20	1.10	1.0	1.09					0.7	1.087
		1.0	1.05					0.7	1.052
50	1.03	1.0	1.03					0.5	1.032
								0.5	1.029
								0.5	1.026
20	1.01							0.5	1.009
								0.4	0.991
20	0.987							0.3	0.986
								0.5	0.984
								0.5	0.981

General Conclusions

Finally it remains to consider the origin of the two phosphates just described. On the basis of the meagre evidence presented by the specimens, it is apparent that the mineral association found within these pockets could have two modes of origin. Firstly the minerals (i.e. whitlockite, woodhouseite and quartz) could be products of the last stages of crystallisation within the apatite veins. They would therefore represent the final impure residues enriched in Al, Si, Mg, Fe, Mn, Na, V, OH, SO_4 etc. after the crystallisation of the relatively pure host chlorapatite.

A second possibility is that these minerals were formed during a subsequent stage of metamorphism, metasomatism, and hydrothermal activity which might have affected the large apatite veins; (for example during the period of granitic-pegmatite introduction which obviously post-dated the apatite veins in this area).

Acknowledgements

Sincere thanks are expressed to Professor T. F. W. Barth and Dr. H. Neumann for their constant advice during the course of this work. The author is also indebted to Professor I. Oftedal for the spectrographic analysis published in this paper, to Miss J. B. Wiik for the x-ray powder diagrams and to Miss B. Mauritz who took the photographs.

LITERATURE CITED

Bannister, F. A., 1947. Whitlockite from Sebdou, Oran, Algeria. Min. Mag., vol. 28, pp. 29–30.

Bredig, M. A., Franck H. H. and Füldner, H., 1932. Beiträge zur Kenntnis der Kalk-Phosphorsäure-Verbindungen. Ztschr. Elektrochem., Bd. 38, Nr. 3.

— 1933, ibid. Bd. 39, p. 959.

Fisher, D. J., 1958. Pegmatite phosphates and their problems. Am. Min., vol. 43, nos. 3 and 4, pp. 181–207.

Frondel, C., 1943. Mineralogy of the Calcium phosphates in Insular phosphate rock. Am. Min., vol. 28, no. 4, pp. 215–231.

Julien, 1865. Am. Jour. Sci., 40, 371. (ref. from Dana).

Kloos, J. H., 1888. Untersuchungen über Gesteine und Mineralien aus West-Indien. Sammlg. d. geol. Reichmuseums. Leiden., Ser. II, Bd. 1.

LEMMON, D. M., 1937. Woodhouseite, a new mineral of the Beudantite group. Am. Min., vol. 22, no. 9, pp. 939—937.

MCCONNELL, D., 1942. X-ray data on several phosphates. Am. Jour. Sci., 240, pp. 649—657.

NEUMANN, H., JØSANG, O. and MORTON R. D., 1960. Mineral occurrences in Southern Norway. Guide to excursions A15 and C12. Int. Geol. Congress, XXI Session, Norden.

PALACHE C., BERMAN, H. and FRONDEL, C., 1951. The system of mineralogy of James Dwight Dana and Edward Salisbury Dana. 7th. ed., vol. II.

STRUNZ, H., 1957. Mineralogische Tabellen, 3rd ed., Leipzig.

TRÖMEL, G., 1932. Mitt. Kaiser Wilhelm Inst. Eisenforsch., 14, 25. (ref. Dana).

TRÖMEL, G. and MÖLLER, H., 1932. Zeits. anorg. Chem., 206, 227. (ref. Dana).

YGBERG, E. R., 1945. Svanbergite from Horrsjørberg. Arkiv før Kemi, Mineralogi och Geologi, 20A, no. 4, pp. 1—17.

Manuscript received July 8, 1960.

Printed December 1961.

PLATE 1.

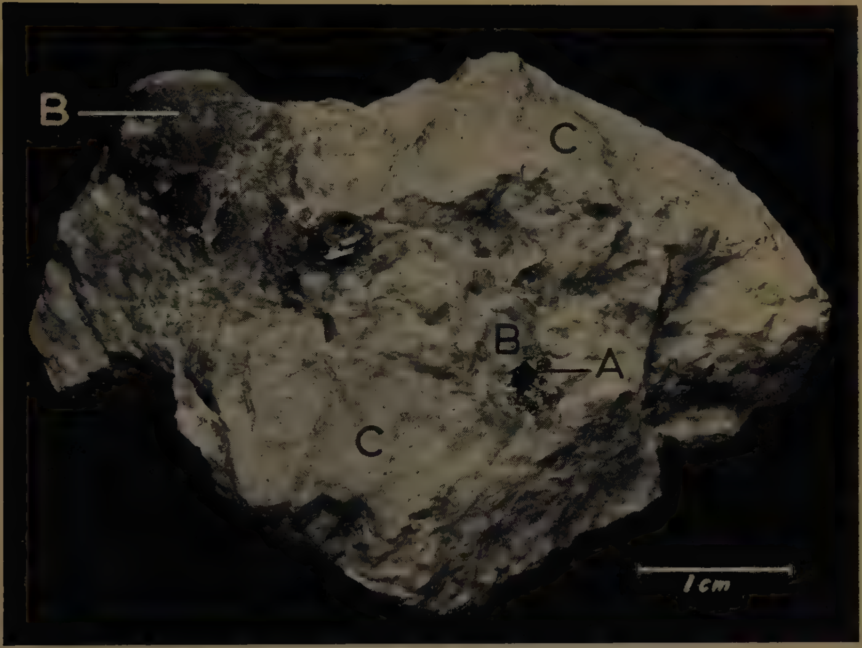


Fig. 1. Specimen from the Ødegården Apatite mines showing massive apatite enclosing pockets containing whitlockite and woodhouseite. A = woodhouseite; B = whitlockite; C = apatite.

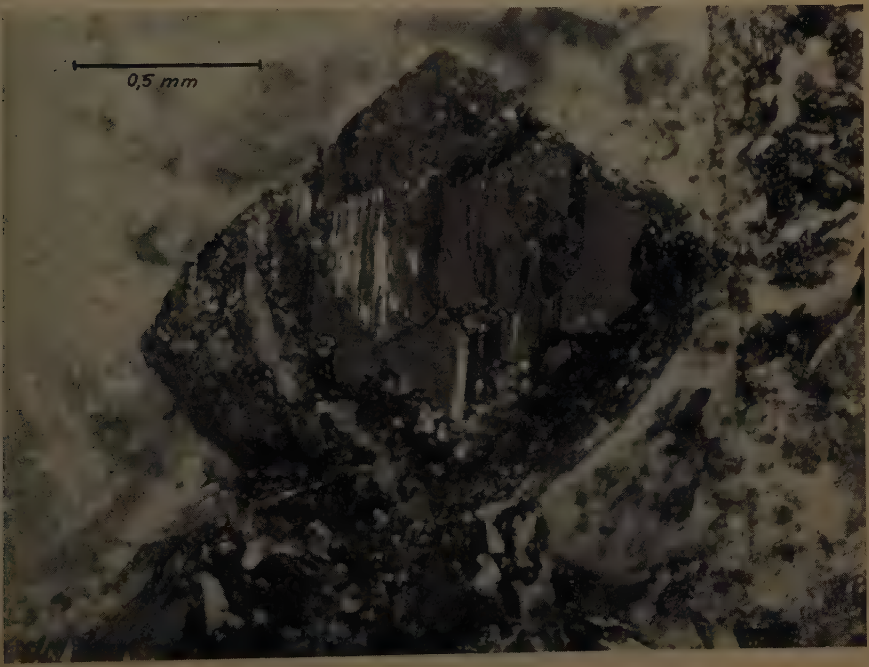


Fig. 2. A euhedral crystal of woodhouseite from the Ødegården Apatite mines illustrating the characteristic form and the nature of the striations seen on the crystal-faces

CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 10. On synchisite in Norway

By

PER CHR. SÆBØ and HENRICH NEUMANN

Abstract. Synchisite is described from Kongsberg, Norway. X-ray powder data are reported. Optical and physical constants are given. The synchisite from Kongsberg shows some differences from the synchisite from Narsarsuk, Greenland.

Introduction

During investigations of anatase in Norway (SÆBØ and DIVLJAN, 1960) the anatase material from Kongsberg (NEUMANN, 1944) was reexamined. These investigations revealed the presence of a hexagonal mineral, which was previously unreported. X-ray work and optical measurements done at the X-ray laboratory, Mineralogisk Geologisk Museum prove that the mineral is synchisite, $CeFCO_3.CaCO_3$. This mineral has previously not been found in Norway.

Synchisite and its paragenesis

During his investigations of the famous native silver deposits at Kongsberg, southern Norway, one of us (H.N.) came across one old specimen from these mines, labelled "Kristians søndre stoll, 1856". This specimen contains anatase in small brilliant crystals, together with albite, quartz, chlorite, small cubes of pyrite, and synchisite. The sequence of the minerals are quartz and albite, then anatase rather simultaneously or in some cases slightly earlier than synchisite, then

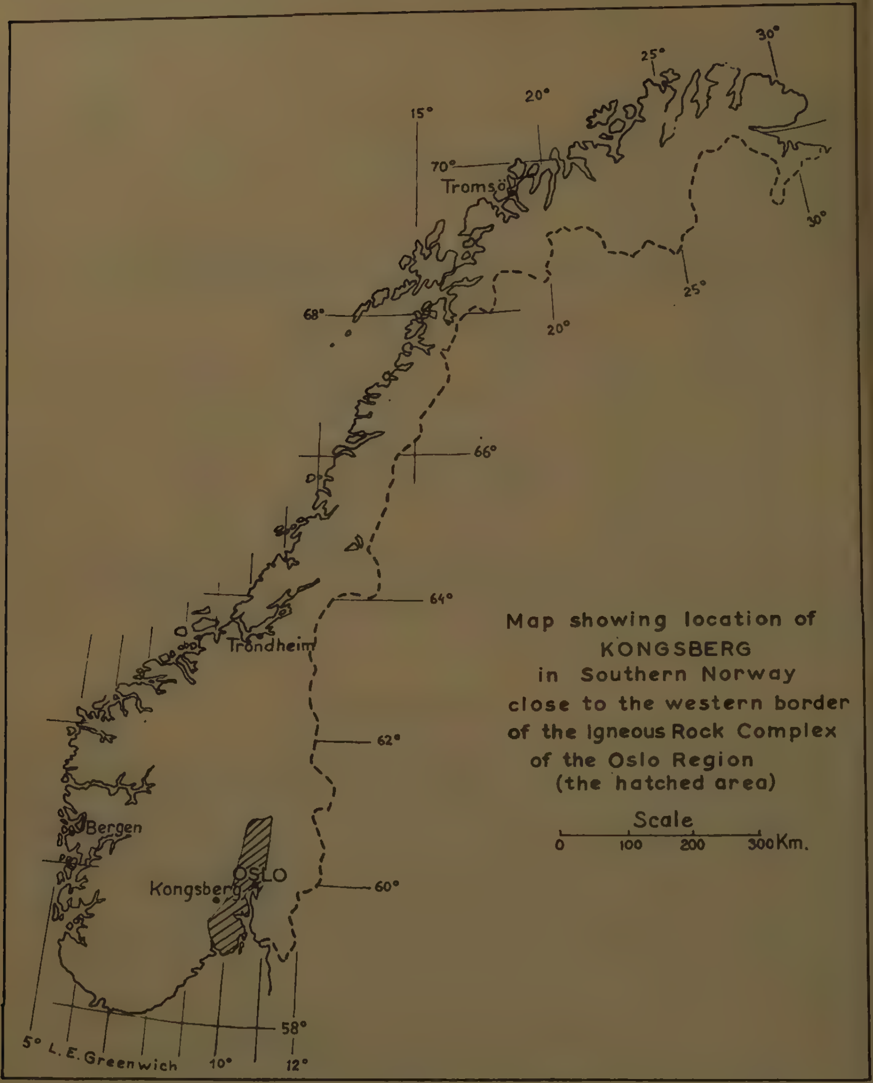


Fig. 1. Situation map.

pyrite and finally chlorite (NEUMANN, 1944). Synchisite occurs as very small, light yellow, hexagonal prisms, 0.2 mm long, slightly modified by small pyramidal faces. The base is well developed and brilliant, the other faces being dull and slightly altered.

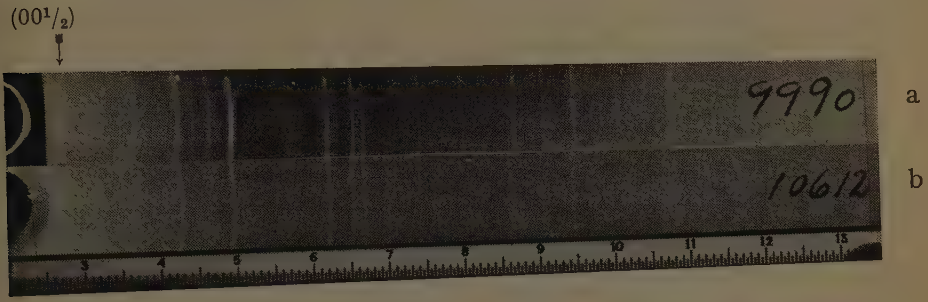


Fig. 2. Reproduction of the X-ray powder patterns. a. Synchisite from Kongsberg. Note the presence of the (00½) reflection. X-ray film no. 9990. Geologisk Mineralogisk Museum, University of Oslo. b. Synchisite from Narsarsuk, Greenland. (00½) reflection is absent. X-ray film no. 10612. Geologisk Mineralogisk Museum, University of Oslo.

The mineral is slowly dissolved in strong HCl with weak effervescence. The crystals are rather brittle, showing no preferred cleavage; fracture conchoidal.

The X-ray powder pattern shows clearly that the mineral belongs to the bastnäsite-synchisite group of minerals. In order to get the exact identification, the powder pattern was indexed on the basis of hexagonal symmetry; and the lengths of the a' and the c''' of the pseudo-cell, as defined by DONNAY and DONNAY (1953), were calculated.

The values of the crystallographic constants and the axial ratio show that the mineral is synchisite. The data of the powder pattern are given in tables I and II.

Small variations in the unit-cell dimensions and physical properties of the members of the bastnäsite-synchisite group are often found (DONNAY and DONNAY, 1953; JANSEN et al., 1959; SVERDRUP et al., 1959). These variations are partly dependent on the (OH:F) — ratio in the minerals. As a rule, however, the OH content is low.

The X-ray powder pattern of synchisite from Kongsberg differs slightly from that of synchisite from Narsarsuk by the presence in the former of a rather strong reflection with d = 9.03 Å, which is absent in the latter. Otherwise the films are identical apart from spacing variations caused by a small difference in the size of the pseudo-cells. The indices of the 9.03 Å reflection are (00½) when indexed on the basis of a pseudo-cell with a_0 = a' and c_0 = c'''. It seems, therefore, that the structure of the synchisite from Kongsberg as different from other synchisites is characterized by a pseudo-cell with twice the normal length of the c-axis, namely with a = a' = 4.07 Å and c = 2c''' = 9.08 Å, which is also the morphological cell of synchisite as stated by DONNAY and DONNAY (1953).

X-ray powder patterns of the synchisite from Kongsberg and Narsarsuk are reproduced as fig. 2.

Optical data

Several immersion slides were studied under the microscope. The mineral is strictly uniaxial (+); any possible cleavage is not observed. The following values for the indices of refraction were found.

$\omega > 1.653$ for yellow light; $\omega = 1.653 \pm 0.005$ for blue light.

$\varepsilon = 1.745 \pm 0.005$; $\varepsilon - \omega = 0.084$.

The exact value of ε is difficult to obtain because of lack of properly oriented grains in the slides.

Genesis of the synchisite

The old label which is glued on to the synchisite specimen gives the only available information about it, namely that it was found in the southern part of the "Kristian adit" in the year 1856. The "Kristian adit" is the main transport adit for the whole complex of silver mines at Kongsberg, running through and sub-parallel to the foliation of the banded gneisses which are the bedrock of the argentiferous veins. The adit passes through a number of the silver mines, but the specimen is hardly found in any of these as in that case the name of the mine would most certainly have been given on the label and not the name of the adit. The specimen is a small slab of rock 3 cm × 6 cm × 10 cm, covered with minerals on three sides, and has characteristically an "upper side" and a "lower side", one of them being covered with scores of crystals of anatase and quartz while the other is not. In all probability this specimen has nothing to do with the native silver veins: it has no calcite which is the dominating gangue mineral of those veins, nor any of the characteristic silver vein minerals, and last but not least, it lacks that undefineable but nevertheless typical general appearance of the vein specimens. One of us has earlier tentatively

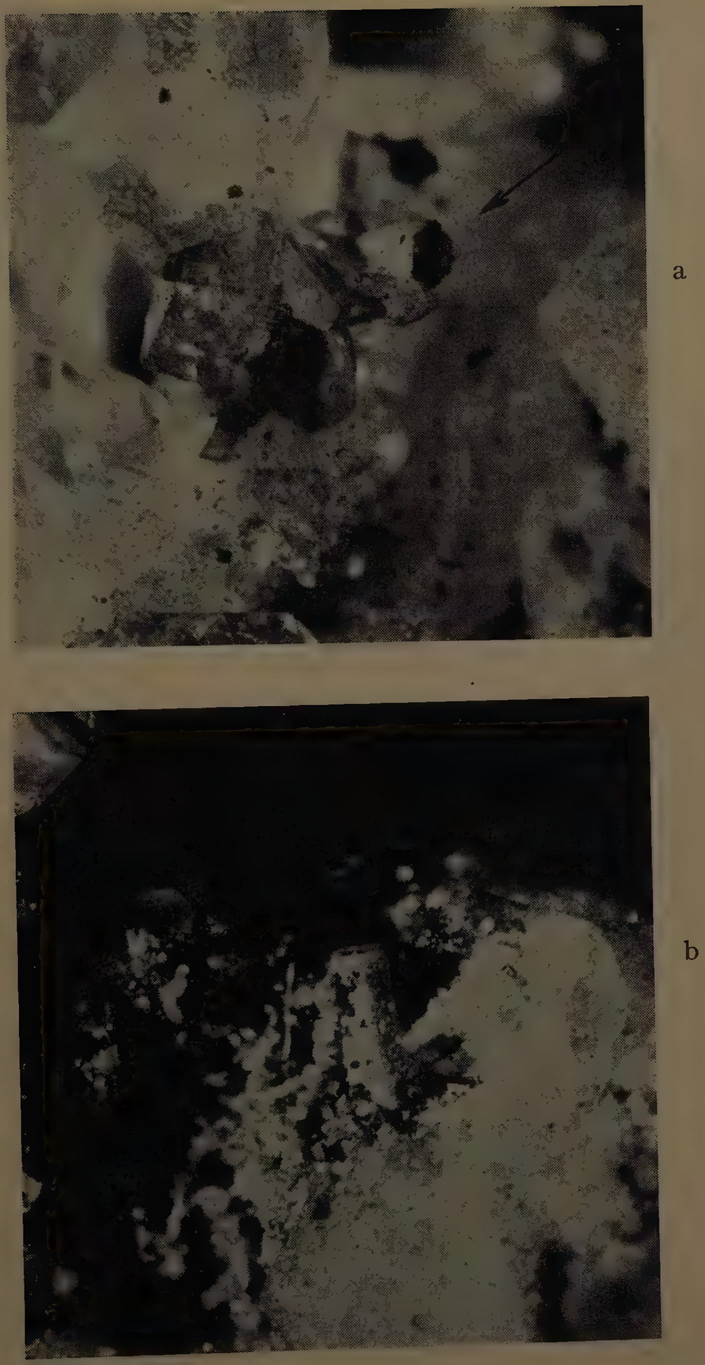


Fig. 3. Photos showing the synchisite from Kongsberg. a. Synchisite crystal (marked by an arrow) together with several anatase crystals. b. Synchisite crystal on quartz. The synchisite is partly overgrown by chlorite.

listed anatase amongst the minerals of the native silver veins (NEUMANN, 1944), this is hardly correct, neither anatase nor synchisite seem to belong to that mineral assemblage.

Most probably the albite-quartz-anatase-synchisite specimen originates from a cavity in the banded gneisses, and its formation may be more or less closely related to that of similar parageneses in alpine veins.

Acknowledgements

We are indebted to Professor I. Oftedal for valuable discussions during this work, and to Dr. P. H. Reitan for correcting the text. Miss D. Engelsrud has prepared the map, and M. Brynhildsrud has taken the photographs.

Table 1.

Synchisite, Kongsberg, Norway. Cam-diam. 9 cm Fe-rad.
a' = 4.07 ± 0.01 Å, c''' = 4.54 ± 0.01 Å, c'''/a' = 1.115

Nr.	Intensity	(hkl)*	θ	d(Å)
1	M		6.16	9.03
2	M	(001)	12.41	4.51
3	V.V.W		14.46	3.88
4	S	(100)	15.99	3.52
5	M		17.20	3.28
6	M		18.60	3.04
7	V.V.S	(101)	20.46	2.77
8	W		25.09	2.29
9	W		25.40	2.26
10	V.S	(110)	28.53	2.03
11	V.W.		29.33	1.977
12	S	(102)	30.67	1.899
13	S	(111)	31.56	1.848
14	W	(200)	33.48	1.756
15	V.W		34.15	1.725
16	V.W.		35.14	1.683
17	M	(201)	36.99	1.640
18	M	(112)	39.92	1.510
19	M	(202)	44.33	1.386
20	W		46.66	1.332
21	M	(211)	49.34	1.277
22	W+		53.10	1.211
23	W		55.53	1.175
24	S	(212)	57.54	1.148
25	W		58.34	1.138
26	W+		63.83	1.079
27	W+		67.95	1.045
28	M		75.32	1.001

V.S = very strong, S = strong, M = medium, W = weak. Intensities estimated visually.

* Hexagonal indices.

Table 2.
X-ray data for synchisite.

a' (Å)	c''' (Å)	c'''/a'	Locality	Author
4.10	4.57	1.114	Narsarsuk	Oftedal (1930)
4.10_3	4.560	1.111	—»—	Donnay and Donnay (1953)
4.07	4.54	1.115	Kongsberg	Sæbö & Neumann 1960

LITTERATURE

DONNAY, C. and DONNAY, J. D. H. 1953. The crystallography of Bastnaesite, Parisite, Roentgenite and Synchisite. Am. Min. Vol. 38, p. 932—963.

JANSEN, GEORGE J., MAGIN JR. GEORGE B. and LEVIN, BETSY. 1959. Synthesis of Bastnaesite. Am. Min. Vol. 44, p. 180—181.

NEUMANN, H. 1944. Silver deposits at Kongsberg. Norges Geol. Unders. Nr. 162, p. 81—82.

OFTEDAL, I. 1931. Über Parisit, Synchysit und Kordylit. Z. Krist. Band 79, p. 437—464.

SVERDRUP, THOR L., BRYN, KNUT ØRN and SÆBØ, PER CHR. 1959. Bastnaesite, a new mineral for Norway. Norsk Geol. Tidsskr., vol. 39, p. 237—247.

SÆBØ, P. CHR. and DIVLJAN S., 1960. On anatase in Norway. In preparation.

Manuscript received November 10, 1960

Printed December 1961.

NORSK GEOLOGISK TIDSSKRIFT

ON SUB-SOLIDUS RELATIONS OF SILICATES

I. The Lower Breakdown Temperature of the Åkermanite Gehlenite Mixed Crystal Series at Moderate Water Pressure

By

OLAV H. J. CHRISTIE
(Mineralogisk-Geologisk Museum, Oslo NÖ)

Abstract. Some sub solidus relations in systems of åkermanite gehlenite compositions in the range 200° C — 700° C have been studied. Åkermanite was reported to decompose to wollastonite and monticellite at approximately 700° C (HARKER and TUTTLE, 1956). At 6 kbar water pressure and 500° C, wollastonite and clinopyroxene form from åkermanite. At the same pressure and ca. 450 °C. clinopyroxene and xonotlite seem to be the stable mineral association. The decomposition temperature of the åkermanite-gehlenite mixed crystal series at 4.8 bar to 6.7 kbar depends upon the composition of the mixed crystal and is situated between 500° C and 650° C. The decomposition products are vesuvianite and hydrogarnet, hydrogarnet, or hydrogarnet and xonotlite in addition to clinopyroxene and wollastonite formed from the åkermanite-rich batches. The hydrogarnets formed during the experiments are mixed crystals in the series hydrogrossular - hydropyrope.

Previous work

Åkermanite ($Ca_2MgSi_2O_7$) and gehlenite ($Ca_2AlSiAlO_7$) are the two most important end members of the melilite group.

Åkermanite was described by J. H. L. VOGT (1884), who found it as an artificial mineral in slags. ZAMBONINI (1910) described natural åkermanite from Vesuvius, and FERGUSON and BUDDINGTON (1919) determined the chemical composition of it.

Gehlenite was described by FUCHS (1815). It was subject to closer study by RANKIN and WRIGHT (1915). The composition of natural melilites was studied by WINCHELL (1924), by BERMAN (1929), and by GOLDSMITH (1947 and 1948).

The minerals of the system åkermanite-gehlenite, in this paper referred to as melilite, form a series of mixed crystals with a minimum melting point at 74 per cent åkermanite (FERGUSON and BUDDINGTON, 1919; OSBORN and SCHAIRER, 1941). This explains the frequent occurrence of zoning in both natural and synthetic melilites.[1]

The breakdown of åkermanite at atmospheric pressure in certain temperature ranges has been reported by several authors. CARSTENS and KRISTOFFERSEN (1931) obtained a mixture of åkermanite, diopside, and glass, from åkermanite glass. Similar results are mentioned by BOWEN, SCHAIRER and POSNJAK (1933) and by OSBORN and SCHAIRER (1941). Osborn and Schairer found the lower stability of åkermanite to be 1325° C at atmospheric pressure. Below this temperature diopside formed as a decomposition product.

HARKER and TUTTLE (1956) investigated the lower limit of stability of åkermanite at high pressure. They found the breakdown products to be wollastonite ($CaSiO_3$) and monticellite ($CaMgSiO_4$) in the range 30—60 kpsi water pressure. These results were extrapolated to zero pressure under the assumption that no other products than wollastonite and monticellite formed from åkermanite.

The lower stability limit of gehlenite under atmospheric pressure was studied by TAVASCI (1939) and by MALQUORI and CIRILLI (1943) with special application to reactions in hydraulic cements. No attempt was made to transfer the results to geological conditions.

Technique and results of the present experiments

The present study is an investigation of the lower stability temperatures of melilite at moderate water pressure. The starting materials for the experiments were prepared by heating the oxides at 1300° C with repeated intermediate grindings until a homogeneous material was obtained. Cold seal bombs, or Morey bombs, were used for the heating experiments. The water pressure was supplied by a manual high pressure pump of conventional design. The pressure readings

---

[1] See following note on the chemistry of melilites.

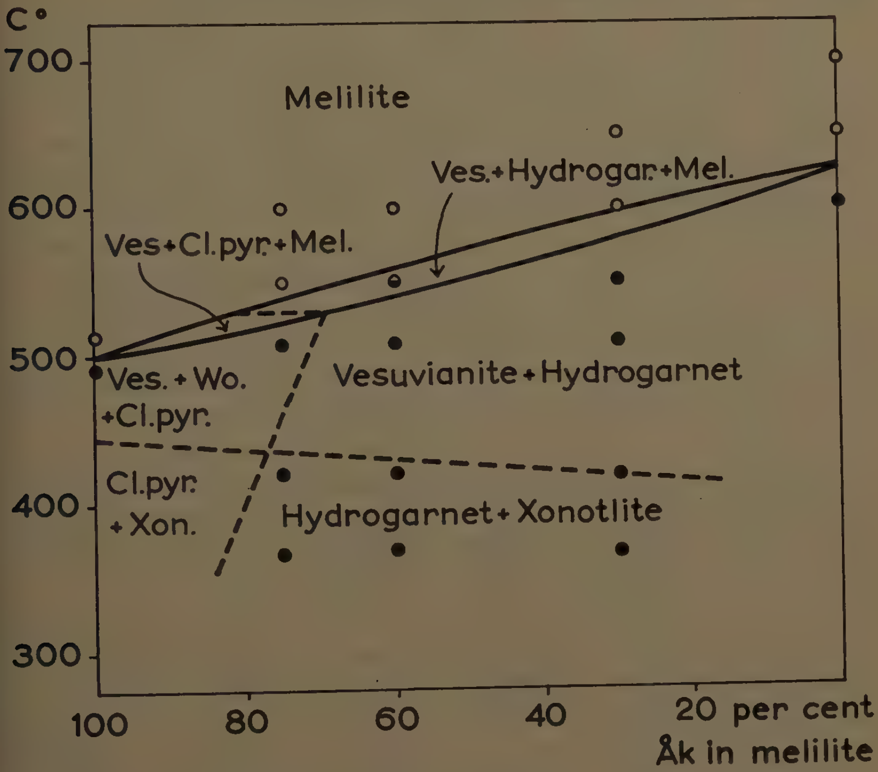


Fig. 1. The low temperature breakdown relations of sodium free åkermanite-gehlenite mixed crystals at 4.8 kbar to 6.7 kbar water pressure. Data from Table 1.

have an accuracy of plus minus ten per cent. The temperature of the furnaces was controlled within plus minus five degrees centigrade by controllers of the falling bridge type. The products were examined by microscope and by nine-centimeter X-ray powder cameras.

There is always a possibility that metastable phases may develop in mineral synthetic experiments. To determine a stability limit both decomposed products and undecomposed ones are heated at identical conditions and if they yield the same end products, this is taken as a proof that true equilibrium is obtained. In the present writer's opinion, however, true equilibrium cannot be stated by thermal experiments, true equilibrium can only be determined by energy calculations.

Table 1.

Products formed by decomposition from åkermanite gehlenite mixed crystals during heating experiments.

Per cent åkermanite in melilite	Water pressure kbars.	Temperature °C	Time	Products
100	4.5	500	9 days	Mel + cl.pyr + wo
100	4.8	550	9 days	Mel
75	6.0	650	11 days	Mel
75	5.5	600	8 days	Mel
75	5.1	550	12 days	Mel
75	6.7	509	10 weeks	Ves + cl.pyr
75	6.5	447	8 weeks	Gar + xon
75	6.5	369	14 weeks	Gar + xon
60	5.5	600	8 days	Mel
60	5.1	550	9 days	Mel + ves + tr. gar
60	6.7	509	10 weeks	Ves + gar
60	6.5	447	8 weeks	Gar + xon
60	6.5	369	14 weeks	Gar + xon
30	6.0	650	10 days	Mel
30	5.5	600	8 days	Mel
30	5.1	550	10 weeks	Mel
30	6.7	509	10 weeks	Gar + ves
30	6.5	447	8 weeks	Gar + xon
30	6.5	369	14 weeks	Gar + xon + tr. bruc
0	6.3	700	6 days	Mel
0	6.0	650	9 days	Mel
0	5.5	600	9 days	Mel + tr. unident. prod.

Thermal experiments can under favourable conditions give us an idea of which phase is the stable one at given conditions. Nevertheless metastable assemblages exist for geological time, they exist in nature more as a rule than as an exception.

In this study all the runs between 600° C and 500° C were made with synthetic melilites as well as with unreacted batches. They yielded identical end products.

The chief decomposition products formed from melilite during the experiments were hydrogarnet, vesuvianite, and xonotlite.

Hydrogarnet ($Ca_3Al_2(SiO_4, (OH)_4)_3$ was first described by Pabst

(1937), later by BELYANKIN and PETROV (1941) and by HUTTON (1943). YODER (1950) studied the formation of synthetic hydrogrossular in his work on the system grossularite — calcium aluminium hexahydrate and found that there is experimental evidence for a complete replacement of SiO_4 groups by $(OH)_4$ groups in grossularite. However, not more than a third of the SiO_4 groups has been replaced by $(OH)_4$ groups in the natural hydrogarnets described by Pabst, by Belyankin and Petrov, and by Hutton.

The natural hydrogarnets described are all hydrogrossular; no hydropyrope has been reported either from natural occurrences or from synthetic products. The present experiments show the existence of a solid solution series in the system hydrogrossular — hydropyrope.

The variation of the unit cell in the hypothetical hydrogrossular — hydropyrope series is presented in the tentative diagram of Fig. 2. Provided the plane of unit cell dimensions of Fig. 2 is not curved the length of the unit cell edge of magnesium aluminium hexahydrate (hydropyrope with complete $(OH)_4$ substitution) is 12.1 Å.

PISTORIUS and KENNEDY (1960) showed that if sufficient water is present to form hydrogarnet an additional increase of the water pressure will not increase the $(OH)_4$ substitution to any appreciable amount; a variation of the temperature, however, will give rise to a distinct change in the $(OH)_4$ substitution.

If pyrope behaves like grossularite any mixed crystal in the hydrogrossular — hydropyrope series will have the same $(OH)_4$ substitution provided the temperature of formation is constant. In that case the a_0 of the unit cell for mixed crystals in the series hydrogrossular — hydropyrope formed at constant temperature will be situated along a line parallel to the constant $(OH)_4$ lines of the plane of Fig. 2.

Fig. 3 displays the variation in the edge of the unit cell, a_0, of hydrogarnets formed by decomposition of melilite. The unit cell edge of pure hydrogrossular is derived from the data of PISTORIUS and KENNEDY (1960).

The lines connecting unit cell edges of hydrogarnets formed at constant temperature have the same slope as the lines of constant $(OH)_4$ substitution of Fig. 2. The slope is derived directly from Fig. 2 and the lines are not drawn to make the best possible conformity with the observed values. It is evident from Fig. 3 that there is a good

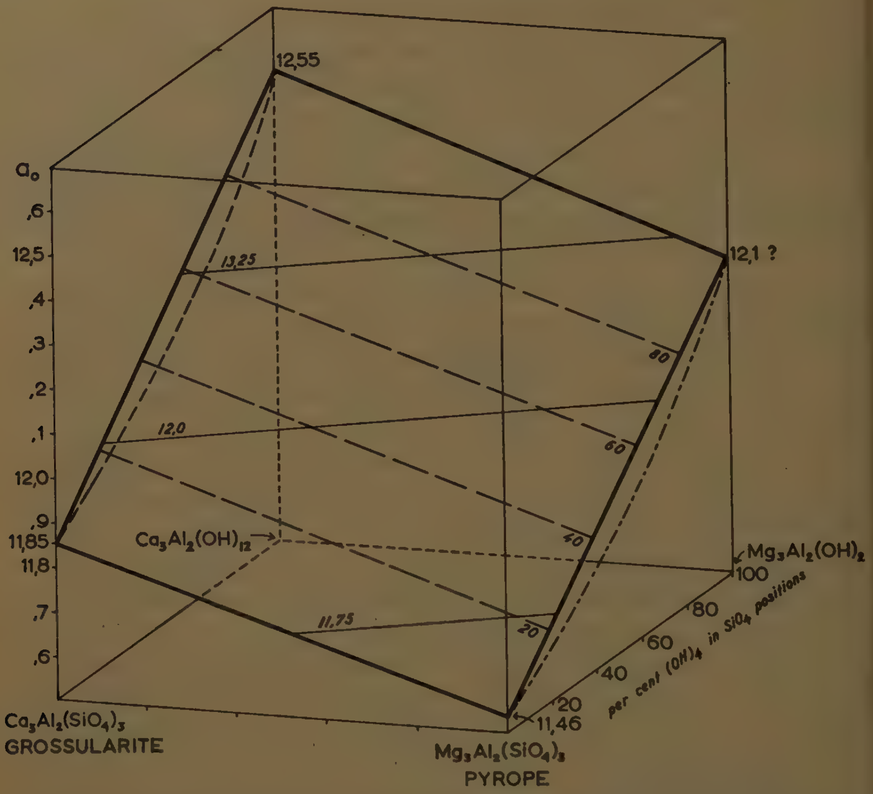


Fig. 2. Length of unit cell edge, a_0, of grossularite-pyrope hydrogarnets. The uncurved plane corresponds to Vegard's law. Dashed line in the grossularite-hydrogrossular plane is taken from Yoder (1950). Dash-and-dot line in the pyrope-hydropyrope plane are estimated values for the hypotetical hydropyrope series. At 10 kbars water pressure hydrogrossular takes no more than 6 per cent hydropyrope in solid solution (Yoder, personal communication 1960). At lower pressure, however, the mutual solubility of the hydrogarnets increases but no quantitative data are available at present. This diagram indicates only the a_0 of possible hydrogrossular-hydropyrope mixed crystals and should not be taken as a suggestion that complete solid solution exists in this system.

agreement between the values predicted from Fig. 2 and the observed values presented in Fig. 3.

The hydrogarnets formed at 369° C have a smaller unit cell than those formed at 447° C. This is probably due to a smaller content of calcium and may be explained by the formation of xonotlite in the low

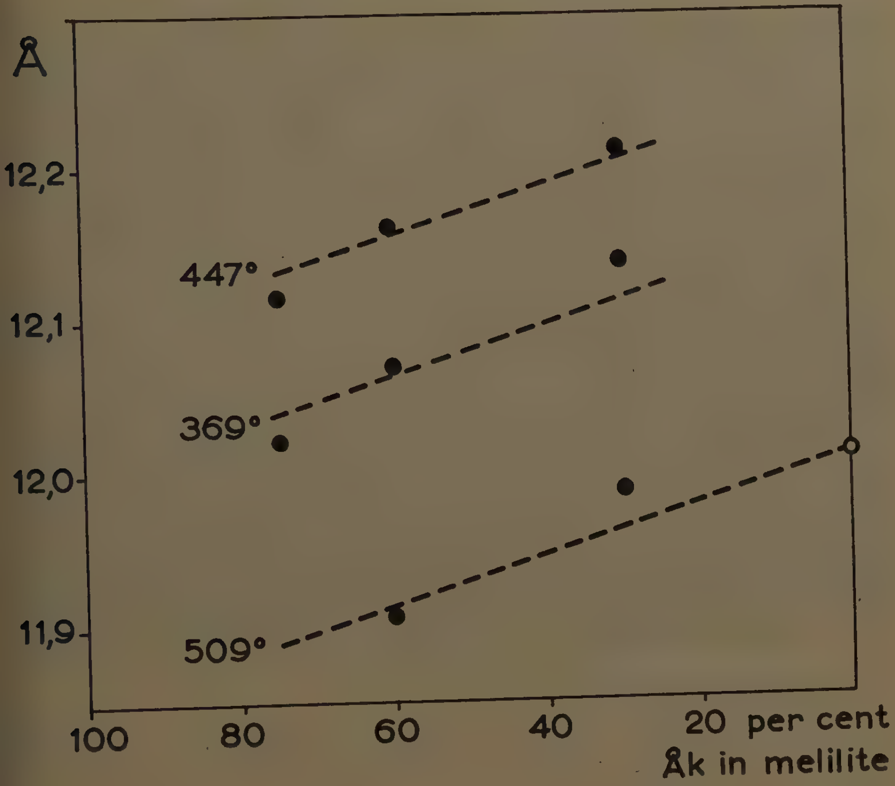


Fig. 3. Unit cell edge, a_0, for hydrogarnets formed by decomposition of melilite. Dashed lines indicate slope of lines of constant $(OH)_4$ substitution of fig. 2. O is taken from Pistorius and Kennedy (1960).

Table 2.

Unit cell size of hydrogarnets formed by decomposition of melilite.

Per cent Åk in melilite	Temperature °C	Unit cell size
75	369	12.02
75	447	12.12
60	369	12.07
60	447	12.16
60	509	11.91
30	369	12.14
30	447	12.21
30	509	11.98

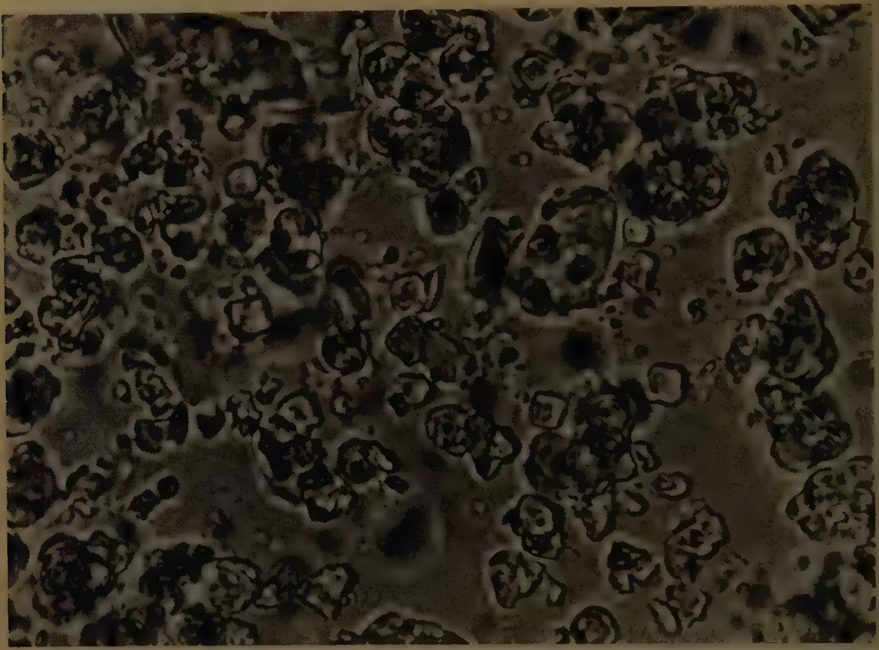


Fig. 4. Hydrogarnets from the system grossularite-pyrope-water formed by decomposition of 30 Åk 60 Geh melilite at 447° C and 6.5 kbar water pressure for 8 weeks. The photo displays 0.1 mm across.

temperature range. The xonotlite requires much calcium, thus making less calcium available for the hydrogarnets.

In one of the runs of 30Åk70Geh at 369° C small amounts of brucite were formed. This might indicate that the hydrogarnet formed at a higher temperature from the same melilite composition is not stable at 369° C.

Vesuvianite ($Ca_{10}Al_4Mg_2Si_9O_{34}(OH)_2$) is well known from contact metamorphic and regional-metamorphic rocks. It has also been found in volcanic ejectamenta.

MACHATSCHKI (1930) investigated the chemical composition of vesuvianites and WARREN and MODELL (1931) determined the structure of it. Both Machatschki and Warren and Modell emphasize the structural, chemical, and physical similarity between garnet and vesuvianite.

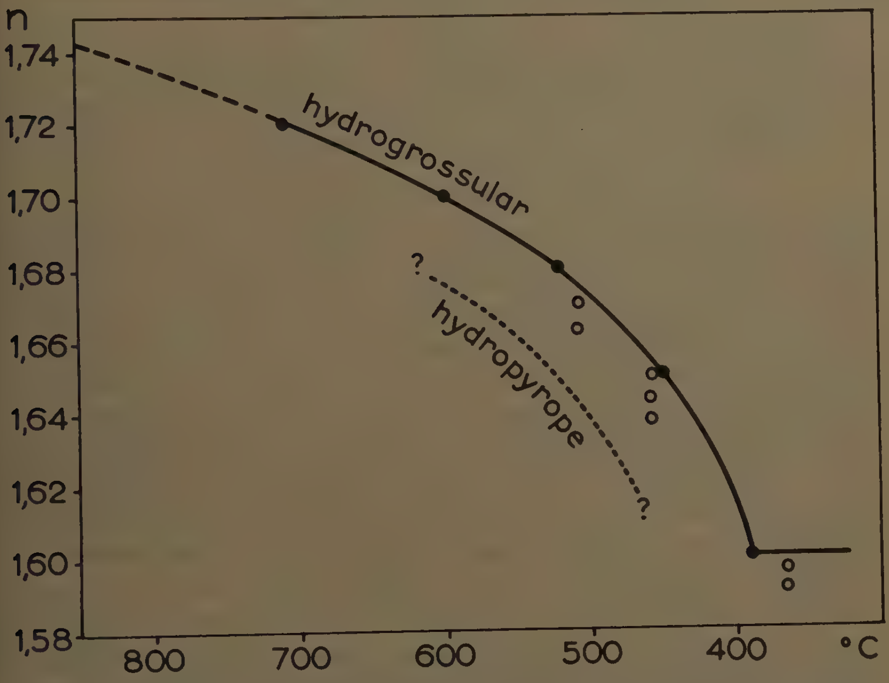


Fig. 5. Refractive index of hydrogarnets formed by decomposition of melilite. Data from table 2 and Yoder (1950).

Vesuvianite is a decomposition product of melilite at 509° C and 550° C. At 447° C and 369° C hydrogarnet and xonotlite, or hydrogarnet alone, formed instead of vesuvianite. The lower critical temperature for the formation of vesuvianite is situated somewhere between 447° C and 509° C. This is confirmed by field experience: vesuvianite occurs mainly in the high temperature contact zones of contact metamorphic rocks.

The refractive index of vesuvianite formed in the present experiments (1.645 — 1.650) is lower than of most natural vesuvianites (1.70 — 1.71). There is also a marked difference in the intensity relation between the line 440, 432, 204, and the line 224, 600, 522 of the X-ray powder pattern of natural and of synthetic vesuvianite. With the present material it has not been possible, however, to show whether or not this phenomenon could be explained from the assumption that vesuvianite, like garnet, is able to form hydro-varieties in which certain SiO_4 groups are replaced by $(OH)_4$ groups.

Table 3.
Refractive index of hydrogarnets formed by decomposition of melilites

Temperature °C	75Åk25Geh	60Åk40Geh	30Åk70Geh
509	—	1.663	1.671
447	1,638	1.642	1.650
369	—	1.595	1.601

Xonotlite ($Ca_6Si_6O_{17}(OH)_2$) never occurred together with vesuvianite during the experiments. Runs at 447° C contained xonotlite, this is above the upper limit of crystallization (390° C) given by FLINT, MCMURDIE and WELLS (1938), but is in good agreement with data recently published by BRUCKNER, ROY and ROY (1960).

The upper stability temperature limit of xonotlite formed in the present experiments decreases with increasing aluminium content of the starting material. This was not expected as the calcium content in mole per cent is constant in all mixtures between åkermanite and gehlenite. However, with increasing aluminium content an increased amount of garnet is formed. Therefore more and more calcium is taken by garnet, and less calcium is left for the residual mixture.

Clinopyroxene ($(Ca, Mg)_2Si_2O_6$) was formed from pure åkermanite at 509° C. HARKER and TUTTLE (1956) reported that åkermanite decomposes at 700° C — 750° C and 2—4 kbar water pressure to wollastonite and monticellite. HARKER and TUTTLE (1956) used heating times up to ten days, whereas the present experiments have been extended several weeks at a higher pressure than those of Harker and Tuttle. This point will be discussed in one of the following papers. The åkermanite-rich batch, 75Åk25Geh, yielded clinopyroxene and wollastonite after ten weeks at 500° C and 6.7 kbar water pressure. This clinopyroxene had the composition 70Di30En (determined by extinction angles). Evidently monticellite (density 3.2) forms at lower pressure than does clinopyroxene (density 3.3). The transition pressure seems to be in the range 4.5 kbar to 5 kbar water pressure.

Wollastonite ($CaSiO_3$) formed from åkermanite together with clinopyroxene. Below ca. 450° C wollastonite is not stable but converts into xonotlite (FLINT, MCMURDIE and WELLS, 1938). Below 450° C the stable association formed from åkermanite is xonotlite and clinopyroxene. If small amounts of gehlenite are present the clinopyroxenes will take aluminium into solid solution. When the clinopyroxene is saturated with aluminium, hydrogarnet will form.

Melilites heated in glycerine

Organic material is supposed to be one of the sources for graphite occurring in schists. Organic material may in many cases accelerate mineral reactions, but it is not known if different compounds have specific catalytic effects upon special mineral reactions.

In the present study some runs were made with synthetic melilites heated in Morey bombs in a glycerine atmosphere. The breakdown products have been identified by X-ray analysis. Examination with the microscope was not successful because of amorphous carbon adhering to the mineral grains.

HARKER and TUTTLE (1956) reported breakdown of åkermanite to calcite and diopside in excess of CO_2 in the range 800° C — 1000° C, and 1.5 — 8 kbar CO_2 pressure. (Estimated values from the diagram of figure 2 in HARKER and TUTTLE (1956)).

There is a breakdown of melilite heated in glycerine in closed Morey bombs at 485° C ± 15° C. The chief decomposition product of these runs was calcite. In one of the runs (300° C) garnet with lattice spacing of pyrope was formed, and in two of the runs boehmite was noticed. Clinopyroxene was not observed in any runs. However, the determination of the decomposition products is not complete, as only X-ray analysis was used for identification.

The decomposition temperature seems to be independent of the Mg–Al ratio of the melilites. As calcite is the only calcium-containing phase detected, except for pyrope, that may have some calcium in solid solution, it is apparently the activity relations of calcium in calcite and in melilite under the given conditions that determine the breakdown temperature of melilite.

Because of the complexity of the interpretation of the physico-chemical conditions of these runs, the experiments have not been carried further. However, they lend support to the idea that organic compounds may accelerate mineral reactions; they may, indeed, act as catalysts by the formation of some minerals typical of the early stages of regional metamorphism. The catalytic effect may be consider-

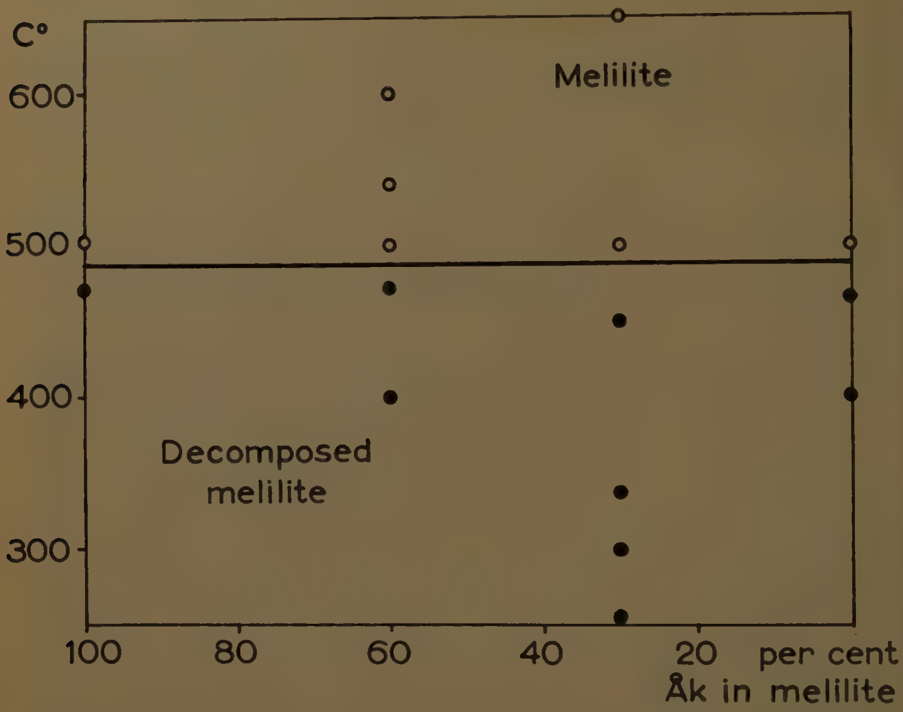


Fig. 6. Decomposition of melilite heated in glycerine in closed Morey bombs.

able, even at low temperatures. Over a period of four days calcite formed from melilite under glycerine pressure at 260° C. Pure gehlenite heated at 200° C for one week showed no presence of calcite, but even at this temperature calcite might form from gehlenite in geological times.

Melilite rocks

Belyankin and Petrov (1941) mention several localities carrying melilite or the minerals formed by the decomposition of melilite: hydrogarnet, and vesuvianite or xonotlite. Mason (1957) described an occurrence of hydrogarnet and melilite in a contact metamorphic limestone. Apart from these descriptions very little is published on natural hydrogarnets formed from melilite-bearing contact rocks.

The melilite dike rocks are better studied. They may be divided into three groups:

Table 4.

Products formed by decomposition from åkermanite gehlenite mixed crystals heated at various temperatures in glycerine atmosphere.

Per cent åkermanite in melite	Temperature °C	Time	Products
100	500	4 days	Mel
100	470	12 days	Cal
60	700	1 day	Mel
60	600	1 day	Mel
60	540	12 days	Mel
60	500	1 day	Mel
60	470	12 days	Cal + Mel
60	400	1 day	Cal + Mal
60	400	12 days	Cal
30	500	4 days	Mel
30	470	12 days	Cal + tr. boehmite
30	400	4 days	Cal
30	340	4 days	Cal + Mel (non equil.)
30	300	4 days	Cal + Garnet
30	260	4 days	Mel + tr. Cal. (non equil.)
30	200	8 days	Mel (non equil. ?)
0	500	1 day	Mel
0	370	12 days	Cal
0	400	4 days	Boehmite + tr. Cal.

1) *Modlibovites* and *Farrisites*, containing melilite, but no monticellite, no garnet and no calcite.

2) *Vesecites, monticellite-Alnøites* and *biotite-Alnøites*, all of them containing melilite and monticellite, but no garnet and no calcite.

3) *Alnøites*, containing melilite, garnet, and calcite among other minerals.

The Modlibovites and Farrisites are probably formed at a higher temperature than than are the other dike rocks mentioned, because none of the melilite decomposition products are present.

Alnöites represent probably low-temperature high-pressure conditions. In the original alnöite (TÖRNEBOHM, 1882) the CO_2 pressure was so high that calcite formed during the crystallization of the melt, as can be seen from thin sections. Here melilite is sometimes intergrown with calcite lamellae in a very fresh rock. The garnets of the alnöites have not been closely studied, and it is not known if they are hydrogarnets.

Vesecites, monticellite-Alnöites and biotite-Alnöites probably represent low-pressure conditions. They contain the low-pressure breakdown product of melilite: monticellite.

None of the melilite dike rocks have been studied with respect to the breakdown products of the melilites. Therefore the physical conditions under which the different melilite rocks formed may only be vaguely suggested. Observations on natural melilite rocks may be explained by help of the results obtained by experiments, and a closer study in light of the present data would be of great interest to petrology.

Acknowledgements

I wish to express my thanks to Norges Teknisk-Naturvitenskapelige Forskningsråd for sponsoring this study and to my father-in-law Mr. Kaare Berntsen, for financial aid during the studies.

LITERATURE

D. S. Belyankin and V. P. Petrov, (1941): The grossularoid group (hibschite and plazolite). Am. Mineralogist *26* (1941) 450.

H. Berman (1929): Composition of the melilite group. Am. Mineralogist *14* (1929) 389.

N. L. Bowen, J. F. Schairer and E. Posnjak (1933): The system CaO-FeO-SiO_2. Am. Jour. Sci. 5th ser. *26* (1933) 193.

D. A. Bruckner, D. M. Roy and R. Roy (1960): Studies in the system CaSiO-$_3$ H_2O. Am. Jour. Sci. *258* (1960) 132.

C. W. Carstens and Kr. Kristoffersen (1931): Schlackenstudien mit besonderer Berücksichtigung der Manganhaltigen Schlacken. N. Jahrb. Min. Beilageband. *62A* (1931) 163.

J. B. Ferguson and A. F. Buddington (1919): The binary system åkermanite gehlenite. Am. Jour. Sci. *50* (1919) 131.

E. P. Flint, H. F. McMurdie and L. S. Wells (1938): Formation of hydrated calcium silicates of elevated temperatures and pressures. — Jour. Res. Nat. Bur. Std. *21* (1938) 617.

J. N. Fuchs (1815): Ueber den Gehlenit, ein neues Mineral aus Tirol. Journal für Chemie und Physik. *15* (1815) 377.

J. R. Goldsmith (1947): The system $GaAl_2Si_2O_8$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$. Jour. Geol. *55* (1947) 386.

— (1948) cited in A. N. Winchell: Elements of optical mineralogy II, 4th Ed. J. Wiley & Sons. p. 473.

R. I. HARKER and O. F. TUTTLE; (1956): The lower limit of stability of åkermanite ($Ca_2MgSi_2O_7$). Am. Jour. Sci. *254* (1956) 468.

C. O. HUTTON (1943): Hydrogrossular, a new mineral of the garnet-hydrogarnet series. Roy. Soc. New Zealand. Trans. Proc. *73* (1943) 174.

F. MACHATSCHKI (1930): Die Summenformel des Vesuvians und seine Beziehungen zum Granat. Centr. bl. Min. A (1930) 284.

G. MALQUORI and V. CIRILLI; (1943): Ric. Sci. Prog. Tecn. Ec. Nac. *14* (1943) 85.

E. F. OSBORN and J. F. SCHAIRER, (1941): The ternary system pseudowollastonite åkermanite gehlenite. Am. Jour. Sci. *239* (1941) 715.

A. PABST (1937): The crystal structure of plazolite. Am. Mineralogist *22* (1937) 861.

C. W. F. T. PISTORIUS and G. C. KENNEDY (1960): Stability relations of grossularite and hydrogrossularite at high temperature and pressure. Am. Jour. Sci. *258* (1960) 247.

F. RAAZ (1930): Über den Feinbau des Gehlenit; ein Beitrag zur Kenntnis der Mehlithe. Akad. Wiss. Wien. Math. nat.w. Kl. Sitzungsber. Abt. 1, *139* (1930) 645.

G. A. RANKIN and F. E. WRIGHT: (1915): The ternary system $CaO\text{-}Al_2O_3\text{-}SiO_2$. Am. Jour. Sci. 4th ser. *39* (1915) 1.

J. V. SMITH (1953): Reexamination of the structure of melilite. Am. Mineralogist *38* (1953) 643.

H. STRUNZ: Mineralogische Tabellen, Ed. Verl. ges. Geest und Portig K. G. Leipzig 1957.

B. TAVASCI: (1939): Struttura del cemento portland idratato. La Chim. et l'Ind. (Milano) *21* (1939) 660.

A. E. TÖRNEBOHM (1882): Mikroskopiska bergartsstudier XVII. Melilitbasalt från Alnö. Geol. För. Sthlm. Förh. VI (1882) 241.

J. H. L. VOGT (1884): Studier over slagger I. Bih. K. Sv. Vet.Akad. *9* (1) (1884).

B. E. WARREN (1930): The structure of melilite $(Ca, Na)_2\ (Mg,Al)\ (Si,Al)_2O_7$. Zeitschr. Krist. *74* (1930) 131.

B. E. WARREN and D. I. MODELL (1931): The structure of vesuvianite $Ca_{10}\ Al_4\ (Mg, Fe)_2\ Si_9\ O_{34}\ (OH)_4$. Zeitschr. Krist. *78* (1931) 422.

A. N. WINCHELL, (1924): The composition of melilite. Am. Jour. Sci. 5th Ser. *8* (1924) 375.

H. S. YODER (1950): Stability relations of grossularite. Jour. Geol. *58* (1950) 221.

F. ZAMBONINI (1910): Mineralogia Vesuviana, Mem. R. Accad. d. Sc. fis. e. mat. Napoli (2) *14* no 7, 1909.

Manuscript received June 22, 1960.

Printed December 1961.

REMARKS ON THE VARIABLE CONTENTS OF Ba AND Sr IN MICROCLINE FROM A SINGLE PEGMATITE BODY

By

IVAR OFTEDAL

Abstract. Based on spectrochemical determinations of Ba and Sr in about 70 microcline samples collected from roughly 75 m^2 of the surface of a pegmatite body, the distribution of these elements within the pegmatite and within single microcline crystals is discussed. This distribution is very far from uniform, and no sign of subsequent adjustment towards an even distribution is observed. The different crystal chemical characters of Ba and Sr are strikingly apparent. The accuracy of spectrochemical Ba- and Sr-determinations is briefly considered.

Data on the distribution of Ba and Sr within a relatively broad pegmatite band in plagioclase gneiss on the Norwegian Skagerrack coast have been published and discussed earlier (1, 2). A number of new determinations have been added later, and the total number of microcline samples examined is now more than 70. These were collected more or less at random within a selected area of about 75 square meters of the pegmatite surface. The pegmatite proper, apart from enclosed gneiss remnants, consists of smaller and larger crystals of microcline perthite with very subordinate oligoclase. Locally it is rich in quartz, which generally appears to be younger than the microcline.

The results of the spectrochemical determinations of Ba and Sr in the microcline samples are shown in the diagram Fig. 1. The samples have been arranged (from left to right) according to decreasing contents of Ba. The vertical lines connecting the curves for Ba and Sr represent individual samples. The samples poorer in Ba than about 500 p.p.m. nearly always represent very large and light coloured

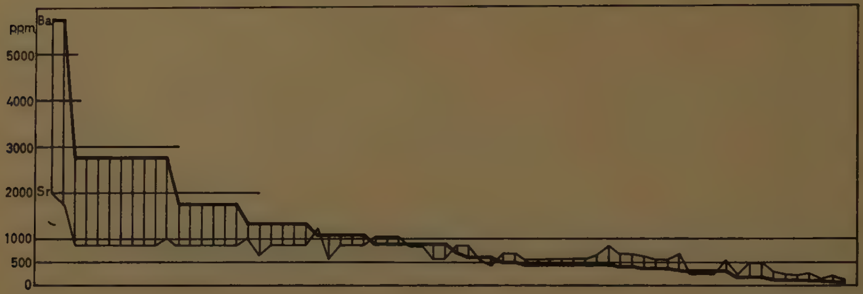


Fig. 1.

crystals, while the others were in general taken from more reddish crystals a few centimeters or less in size. As the large crystals have developed in the interior of the pegmatite, evidently in complete absence of tectonic movements — which have otherwise conspicuously affected the rocks of the area —, they apparently belong to late stages in the development of the pegmatite. Thus it may be supposed that the abscissa axis in the diagram corresponds roughly to a time axis. This is also supported by the fact that the highest concentrations of Ba and Sr plotted in the diagram approach those found in microcline crystallized in actual contact with the plagioclase of the gneiss, as observed in all examined cases of small pegmatite veinlets in the same area. It is obvious that also the crystallization of the larger pegmatite bodies in the area must have started along the contacts towards the country rock. As pointed out earlier (1, p. 236), the distribution of Ba and Sr between the microcline and the adjacent plagioclase in such cases most probably approaches equilibrium, since it is similar to the distribution of Ba and Sr between the feldspar phases of Archaean granites. According to observations on several granites, this distribution is as follows. If the Ba-concentration is x in the plagioclase, it is about 10x in the microcline; then the Sr-concentration is not far from 10x in the plagioclase and x in the microcline. The value of x is about 200 p.p.m. for the Østfold granite and the Birkeland granite. In the present case (contact gneiss-pegmatite) x is about 500 p.p.m., giving a Ba-concentration in the microcline of about 5000 p.p.m., which is in good agreement with the observations. It is seen that the highest Ba-concentrations shown in the diagram are also of this magnitude, though the corresponding Sr-concentrations differ considerably from 500 p.p.m. These microclines evidently approach rather closely the supposed equilibrium with the plagioclase of the gneiss.

It was concluded earlier (1, p. 238) that the pegmatite-forming medium must have been originally very poor in Ba and Sr, that it absorbed Ba and Sr from the plagioclase during its passage through the gneiss, and that the microcline crystals forming from it in direct contact with plagioclase crystals attained approximate equilibrium with these as to Ba- and Sr-contents. It may be of interest to reconsider the development of the large pegmatite in view of the data presented in the diagram. We have seen that the arrangement of the individual samples in the diagram probably corresponds roughly to the succession of crystallization. This being accepted, it is seen that Ba and Sr behave rather differently during the process of development. Apart from irregularities the Sr-concentration decreases much more slowly than the Ba-concentration. For Ba-concentrations lower than about 500 p.p.m. nearly all microclines are considerably richer in Sr than in Ba, and the opposite holds true for Ba-concentrations higher than about 1 000 p.p.m. This demonstrates well the characteristic crystal chemical difference between Ba^{2+} and Sr^{2+} in alkali feldspar crystals: Ba^{2+} is much more readily "captured" than Sr^{2+} (under the existing conditions, which, judging from the mineralogical composition of the surrounding gneiss complex, correspond to low amphibolite facies). This is a good example of preferred diadochy because of close similarity in ionic radii (K^+ 1.33, Ba^{2+} 1.34, Sr^{2+} 1.12). Part of the Sr is probably incorporated in the perthite lamellae (Na^+ 0.97, Ca^{2+} 0.99); individual plagioclase crystals within the pegmatite are much richer in Sr than most of the microcline crystals (usually 1 500 to 3 000 p.p.m.).

The whole succession of observed Ba- and Sr-concentrations in the microcline crystals might be explained on the simple assumption that the pegmatite-forming medium were an actual magma occupying the present site of the pegmatite. Such a magma might have accumulated through narrow veins like the ones now frequently observed as pegmatite veinlets in the gneiss, and subsequently intruded into place. The Ba- and Sr-concentrations in this magma must have been close to the average values calculated from all of the above determinations, which are 1 100 p.p.m. Ba and 700 p.p.m. Sr. If all of this Ba and Sr has been extracted from the plagioclase of the country rock, it is evident that also the *liquid* K-rich medium must have had a much greater affinity for Ba than for Sr, being actually richer in Ba and much poorer in Sr than the adjacent plagioclase. If this magma really existed, it must have had a somewhat higher temperature than the country rock, so that crystallization started at the contacts. The alkali feldspar crystals first formed were highly enriched in Ba as compared with the liquid, showing that Ba had a stronger tendency to occupy the K-sites in the crystals than K itself. As a result the remaining liquid was impoverished in Ba. The next crystals to form must consequently be poorer in Ba, and the remaining liquid still more so. Continuing this process, eventually very Ba-poor crystals would form. The behaviour of Sr is similar, but the enrichment in the crystals relative to the liquid is much less pronounced. It is seen from the diagram that the lowest Ba-concentration observed is about 200 times less than the highest; for Sr (in the same samples) the corresponding figure is about 20. We may say that crystallizing alkali feldspar under these conditions is about ten times more efficient in capturing Ba than in capturing Sr; still Sr is very conspicuously captured. Since Ba and Sr are chemically very similar, it is natural to correlate this ratio of "efficiency" with the radius ratios Ba^{2+}/K^+ and Sr^{2+}/K^+, which are practically unity and 0.84 respectively*. If this "crystallization differentiation" process could be continued further than in this particular instance, the Ba-contents and eventually the Sr-contents of the resulting alkali feldspars would approach zero asymptotically. Several examined amazonites (from other districts) all appear to represent very advanced stages in such a process (1, p. 243). Following this gradual crystallization of the pegmatite, there has been no observable adjustment of the Ba- and Sr-concentrations towards an even distribution within the pegmatite.

Although the hypothesis of a real pegmatite magma thus explains the observed facts very well, this does not of course prove that it is correct. At any rate it represents a simplified picture of the actual conditions, which appear rather complex in the field. If it is essentially correct, it is not necessary to assume that the K-rich pegmatite-for-

---

* Added in proof: According to later analyses by flame photometer, the composition of the microcline is close to Or_{80} Ab_{20} irrespective of varying Ba- and Sr- contents.

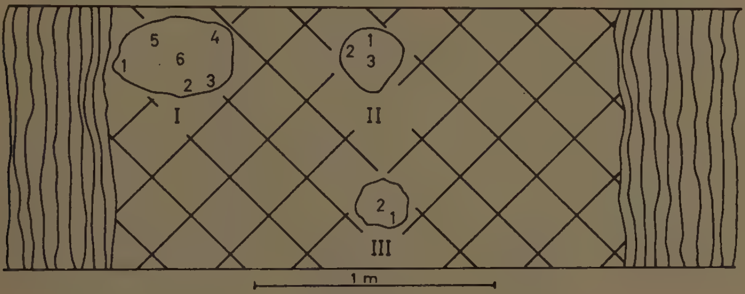


Fig. 2.

ming medium was originally extremely poor in Ba and Sr; it may very well, prior to the intrusion, have contained Ba and Sr in concentrations corresponding to the present averages for the pegmatite as given above.

If the hypothesis of a magma — or a medium behaving more or less like a crystallizing magma — is not accepted, it is difficult to conceive an explanation other than the one offered earlier, viz. that essentially Ba- and Sr-free K-rich fluids were continuously or intermittently introduced from below into the interior of the originally narrow pegmatite band, forming new alkali feldspar crystals which had to be successively poorer in Ba and Sr because the limited supply of these elements available there was gradually used up by the growing crystals. This explanation seems to be more in accordance with the appearance of the pegmatite in the field, but considering the high Ba- and Sr-contents of every examined pegmatite veinlet in the area, it is difficult to imagine a K-rich pegmatitic medium moving for a presumably long time through Sr- and Ba-rich gneiss without acquiring itself a considerable content of these elements. A differentiation process, not necessarily in a magma proper, as outlined above may be more probable after all.

Some observations on 3 large and evidently late microcline crystals in the same pegmatite may be specially mentioned. Fig. 2 shows schematically how these crystals appear on the rock surface. They are embedded in somewhat less coarse pegmatite. The crystals are numbered I, II, III, and the individual samples are indicated by 1, 2, – – – within each crystal. The results of the spectrochemical examination of these samples are shown in Table 1.

Table 1.

Sample no.	Sr, p.p.m	Ba, p.p.m
I, 1	700	500
2	600	450
3	700	500
4	650	250
5	650	300
6	700	400
II, 1	300	150
2	300	150
3	300	150
III, 1	700	350
2	700	600

Before commenting on these results some words on the accuracy of the determinations may be added. Sr and Ba are more difficult to determine quantitatively by spectrochemical analysis than many other elements; parallel exposures may give rather different results. In the present case utmost care was taken to keep the conditions of arcing and development identical throughout, and most of the samples were examined by 2 or more parallel exposures. In addition one of the K-lines was used as an internal intensity standard; only spectrograms showing a certain specified intensity of this K-line were used for the determinations. (It is true that K is not a suitable element for intensity comparison with Ba and Sr, but at any rate the K-line affords a check on the general intensity level of the spectrograms.) A few examples of individual results are shown in Table 2.

Table 2.

Sample no.	Sr, p.p.m.	Ba, p.p.m.
I, 1	650, 650, 800, 650	400, 400, 600, 600
I, 3	650, 800, 650	400, 500, 600
I, 4	650, 650, 650	250, 250, 250

The maximum deviation from the mean is here about 20 %. On the other hand it is seen that the observed deviations may be much smaller than this. More parallel exposures would doubtless narrow the probable limits of error considerably.

In view of these considerations the Sr-concentrations are evidently fairly constant within each of the crystals. The Ba-concentrations, on the other hand, are certainly variable within crystals I and III. Crystal II shows very consistent values, which are much lower than the corresponding ones for crystals I and III. According to the above discussion on the development of the pegmatite, crystal II ought to be "later" than crystals I and III. The distribution of Ba within crystals I and III indicates non-equilibrium, which evidently reflects a variable supply of Ba during the growth of the crystals. Even these very local variations have not been removed by subsequent diffusion. Of course this is not really unexpected in view of the frequent occurrence of e.g. zoned crystals in eruptive rocks. The series K-feldspar-celsian is in fact crystal chemically strictly analogous to the albite-anorthite series. It appears that Ba-ions (Ca-ions) once captured in K-feldspar (albite) are so strongly bound to their sites that they will not migrate in ordinary conditions.

REFERENCES

(1) Ivar Oftedal. Norsk geol. tidsskr., 1958, *38*, 231
(2) — — Norsk geol. tidsskr., 1959, *39*, 343

Manuscript received February 8, 1961

Printed December 1961

THE MIDDLE ORDOVICIAN OF THE OSLO REGION, NORWAY

7. Trilobites of the suborder Cheirurina

By

FRANK NIKOLAISEN

Abstract. Sixteen cheirurid trilobites are described from the Middle Ordovician Ogygiocaris and Chasmops series of the Oslo Region of Norway. The genus *Cyrtometopella*, and also the species *Paraceraurus helgoeyensis*, *Cyrtometopella tumula*, *C. askerensis*, *Nieszkowskia norvegica*, *Hemisphaerocoryphe inflatus*, and *Atractopyge gracilis* are new. The species *Cybele brevicauda*, *C. coronata*, *C. dentata*, *C. kutorgae*, *C. revaliensis*, *C. rex*, and *C. woerthi* are transferred to the genus *Atractopyge*. The species *C. grewingki* (SCHMIDT, 1881) is regarded as a synonym of *Atractopyge dentata* (ESMARK, 1833). The species *Hemisphaerocoryphe? rosenthali* (SCHMIDT, 1881) is transferred to the genus *Cyrtometopella* n. gen.

A chart shows the stratigraphic occurrence of the described species.

CONTENTS

Page
Abstract 279
Introduction 281
Acknowledgements 282
Abbreviations 282
Names used by previous authors 283
Family Cheiruridae 283
  Subfamily Cheirurinae 283
    Paraceraurus helgoeyensis n. sp. 283
      — *perlongus* (BRØGGER) 286
  Subfamily Cyrtometopinae 287
    Genus *Cyrtometopella* n. gen. 287
    Cyrtometopella tumula n. gen., n. sp. 288
      — *askerensis* n. gen., n. sp. 289

Subfamily Acanthoparyphinae 291
  Nieszkowskia norvegica n. sp. 291
Subfamily Deiphoninae 292
  Hemisphaerocoryphe inflatus n. sp. 292
    — *sphaericus* (ESMARK) 293
Family Pliomeridae 295
Subfamily Pliomerinae 295
  Pliomerops sp. 295
Family Encrinuridae 296
Subfamily Encrinurinae 296
  Encrinurus sp. 296
Subfamily Cybelinae 296
  Atractopyge brevicauda (ANGELIN) 296
    — *dentata* (ESMARK) 298
    — *gracilis* n. sp. 302
    — *kutorgae* (SCHMIDT) 304
    — aff. *kutorgae* (SCHMIDT) 305
    — sp. 306
Subfamily Staurocephalinae 306
  Staurocephalus sp. 306
References 307
Middle Ordovician cheirurid trilobites found in Norway and their known stratigraphic and geographic occurrence (chart) 309

Plates 1—4 and explanation of plates.

Introduction

The present paper appears as no. 7 in a series dealing with the Middle Ordovician stratigraphy and fossils of the Oslo region. Team work on this subject was initiated by Professor Leif Størmer in 1950, and is supported by grants from the Norwegian Research Council for Science and Humanities.

The term Middle Ordovician is taken in the sense defined by STØRMER (1953), *i.e.* from and including the Llandvirnian zone of *Didymograptus bifidus* to and including the Caradocian zone of *Dicranograptus clingani*. The Cambro-Silurian of the Oslo region is divided into several districts, cf. STØRMER (1953, p. 51; map p. 53).

The first description of cheirurid trilobites based on Norwegian material was the two Middle Ordovician species *Trilobites sphaericus* (now *Hemisphaerocoryphe sphaericus*), and *Trilobites dentatus* (now *Atractopyge dentata*) described by ESMARK in 1833.

Since then only two species of cheirurid trilobites have been described from the Middle Ordovician of Norway, namely *Cheirurus perlongus* (now *Paraceraurus perlongus*), described by BRØGGER in 1882, and *Staurocephalus* sp., described by KIELAN in 1957.

Only one species previously described from other countries has so far merely been recorded from the Middle Ordovician of the Oslo region (namely *Atractopyge brevicauda*), *e.g.* KIÆR (1897), and HOLTEDAHL (1909).

In 1881 SCHMIDT published his paper «Revision der ostbaltischen silurischen Trilobiten», which contains descriptions of the Baltic species of the three trilobite families Phacopidae, Cheiruridae, and Encrinuridae. This paper has been valuable in determining the Norwegian species of the suborder Cheirurina.

The Norwegian species do not resemble closely any British or North American species except that *Paraceraurus helgoeyensis* n. sp. recalls *Ceraurinus craigensis* (TRIPP, 1954) from the Caradocian mudstones at Craighead Quarry near Girvan, Scotland.

Unfortunately most of the present material consists of detached parts of the shield, whereas complete dorsal shields are rare. Only two nearly complete dorsal shields have been found, and they belong to the same species. For this reason, and because the preservation is not always too good, there are some specimens in the collections which I have not been able to determine. Some of these specimens may belong to other species than those described in this paper. It is regrettable that stratigraphic data are inadequate or even missing for some of the specimens in the collections.

The species *Pliomerops* sp. comes from Redalen in Biri, and thus it should not have been dealt with in this paper, but as this species is the only known Middle Ordovician cheirurid trilobite that occurs in Norway outside the Oslo region, I have decided to include it here.

The stratigraphy used in this paper was published by Størmer in 1953. For abbreviations in this connection see below.

I mainly use the descriptive terms published in the Treatise on Invertebrate Paleontology, Volume O. The terms S1, S2, and S3 used for the lateral glabeller furrows were suggested by Jaanusson in 1956, and Henningsmoen in 1957.

Acknowledgements

I wish to thank Professor Anatol Heintz (director, Paleontological Museum of the University of Oslo) for placing the material at my disposal, for letting me work at the museum, and for placing the necessary equipment at my disposal. I also owe Dr. Gunnar Henningsmoen (Paleontological Museum of the University of Oslo) a debt of gratitude for his valuable help. He has helped me in preparing the fossils, discussing intricate questions with me, and has lent me literature.

I wish to express my sincere thank to Miss. B. Mauritz for her interest in taking good photographs, and to Dr. Frank Vokes for reading the English of the manuscript.

ABBREVIATIONS

The following abbreviations are used in the text:

P.M.O. = Paleontological Museum, University of Oslo, Norway.
Ceph. = Cephalopod Shale (4a*β*?).
Ech. = Echinosphaerites Beds (4b*α*?).
Coel. = Coelosphaeridium Beds (4b*β*?).
Cycl. = Cyclocrinus Beds (4b*γ*?).
Sphaer. = Sphaeronid Limestone (4b*δ*?).

Names used by previous authors

Names used by Esmark, 1833:	Revised names:
Trilobites sphaericus	*Hemisphaerocoryphe sphaericus*
— *dentatus*	*Atractopyge dentata*
Names used by Brøgger, 1882 and 1887:	
1882 *Cheirurus perlongus*	*Paraceraurus perlongus*
1887 *Sphaerocoryphe* cf. *Hübneri*	*Hemisphaerocoryphe inflatus*
1887 *Cybele* sp. (4b*a*)	*Atractopyge dentata*
1887 — sp. (4bδ)	— *gracilis*
Names used by Kiær, 1897:	
Cybele brevicauda	*Atractopyge brevicauda*
Names used by Holtedahl, 1909:	
Cybele brevicauda	*Atractopyge brevicauda*
— *Grewingki*	— *dentata*

Suborder Cheirurina Harrington & Leanza, 1957

[Erroneously ascribed to Öpik, 1937, by Harrington & Leanza, 1957]. [Superfamily Cheiruracea Öpik, 1937; Cheiruroidae Hupé, 1953 (attributed to Öpik, 1937)] [Type — *Cheirurus* Beyrich, 1845]

Family Cheiruridae Salter, 1864

[= Chirurides Hawle & Corda, 1847; Chiruridae Angelin, 1854, suppression of both pend. ICZN; = Cerauridae Miller, 1889]

SUBFAMILY CHEIRURINAE SALTER, 1864

[*nom. transl.* Raymond, 1913 (*ex* Cheiruridae Salter, 1864)]

Genus *Paraceraurus* Männil, 1958

Type species: — *Ceraurus aculeatus* Eichwald, 1860

Paraceraurus helgoeyensis n. sp.

Pl. 1, figs. 1—3, text fig. 1.

Name: — From the island of Helgøy in Lake Mjøsa in the Nes—Hamar district.

Type data: — Holotype (here selected) is a cranidium (P.M.O. no. 67937), from the Mjøsa Limestone (4bδ) at Bergevika, Helgøya.

Material: — In addition to the holotype, 3 more or less fragmentary cranidia (P.M.O. nos. 67915, 67926, 67977 and counterpiece 67978). Furthermore 1 hypostoma (P.M.O. no. 68010), and 3 more or less incomplete pygidia (P.M.O. nos. 67914, 67981 and 69634).

Diagnosis: — Cranidium more than twice as wide as long, glabella with subparallel sides, smooth, and subtrapezoidal in front. Fixigenae wide, pitted, reticulate, and with flattened genal spines pointing obliquely backwards.

Hypostoma with a granulated middle body, and a pair of distinct furrows. Posterior part of the brim wide (long.).

Thorax unknown.

Pygidium flattened, with 2 pairs of pleural spines, the anterior pair much stronger than the posterior pair, and the axial piece with 2 small spines.

P. helgoeyensis n. sp. differs from the type species in having the eyes closer to the glabella, and a wider and more parallel-sided glabella.

Description: — Cephalon subsemicircular in outline, moderately convex transversely. Cranidium only slightly less than 2.3 times as wide as long. Glabella with subparallel sides, and subtrapezoidal in front. Three pairs of strong and distinct lateral glabellar furrows. S1 geniculate, oblique backwards and reaching the occipital furrow. S2 and S3 slightly convex, slightly oblique backwards, reaching one-third across the glabella. S2 slightly less convex than S3. Glabella appears to be smooth, but the test is missing. Occipital furrow deep and distinct, simple and convex. Preoccipital glabellar lobe subrectangular. Occipital ring as wide (tr.) as glabella, mesial tubercle not developed. Axial furrow deep and distinct. Postocular fixigenae slightly less than 1.5 times as wide (tr.) as occipital ring, pitted faint near the glabella, but stronger at the sides. Relatively long, curved and flattened genal spines, pointing obliquely backwards. Posterior border furrow deep and well defined. Posterior border broad (long.). Eyes located at transverse mid-line of the cranidium. Palpebral lobes strong and protuberance-like. Eye ridges slightly oblique backwards, broad (long.) and short (tr.). Width of interocular fixigenae about one-third of that of glabella at eye-line. Preglabellar field narrow (long.). Preocular part of facial suture converging slightly, postocular part running slightly obliquely forwards until reaching the lateral border furrow, and then obliquely backwards. Librigenae small and with

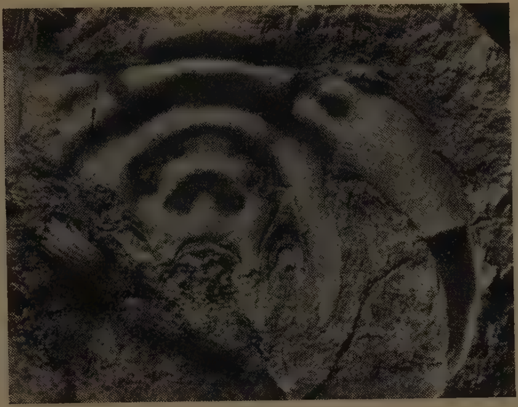


Fig. 1.
Pygidium of *Paraceraurus helgoeyensis* n. sp., from the lower part of the Mjøsa Limestone (4bδ_1), Bergevika, Helgøya, Nes-Hamar. P.M.O. no. 69634. Coll.: F. Nikolaisen, 1961. x 4.0.

well rounded outer margin, lateral border broad (tr.), lateral border furrow deep and distinct.

Hypostoma with a rather convex and granulated middle body. A pair of deep and distinct furrows separates the anterior lobe from the posterior lobe, each furrow reaching about one-third across the middle body. Lateral parts of the brim narrow (tr.), posterior part wide (long.), and with a well rounded posterior margin.

Thorax unknown.

Pygidium flattened, with 2 pairs of pleural spines. The axis is short, occupying about one-fourth of total width. Anterior pair of pleural spines long and broad (tr.), slightly curved, posterior pair about half as long, and half as wide. The endlobe is slightly convex, and provided with 2 pit-like impressions near the base, and a pair of small spines, which seem to be convex in the longitudinal direction, and to be hollow.

Dimensions: — The holotype is 12.1 mm long and 27.5 mm wide. The largest cranidium is 16 mm long and 36.5 mm wide. Largest pygidium is about 7.3 mm long and 22 mm wide (spines not included).

Occurrence: — Mjøsa Limestone (4bδ), NES—HAMAR (Helgøya).

Affinities: — *P. helgoeyensis* n. sp. is closely related to the two Baltic species *P. exsul* (BEYRICH, 1846) and *P. gladiator* (EICHWALD, 1857). It differs from *P. exsul* in having a wider and apparently smooth glabella. Furthermore the eyes are located further forwards. It differs from *P. gladiator* in having the eyes further forwards and closer to the glabella. The pygidium differs from both species in having only 2 pairs of pleural spines.

Furthermore it seems to be related to the species *Ceraurinus craigensis* (TRIPP, 1954), which may be a species of the genus *Paraceraurus*. *P. helgoeyensis* n. sp. differs from this species in having a shorter and wider glabella. The pygidium differs in being narrower.

Paraceraurus perlongus (BRØGGER, 1882)

Pl. 1, fig. 4.

1882 *Cheirurus perlongus* n. sp. — BRØGGER, p. 134, pl. V, fig. 14. (Short diagn. and fig. of glabella).

Type data: — Holotype (by monotypy) is a fragmentary cranidium (P.M.O. no. 2636) probably from the Upper Chasmops Shale (4b*y*) from Oslo. (Figured by Brøgger, 1882).

Material: — The holotype is the only known specimen.

Diagnosis: — A *Paraceraurus* species with subparallel-sided glabella, and 3 pairs of lateral glabellar furrows, all oblique backwards. Occipital ring as wide as glabella. Glabella granulated.

P. perlongus differs from the type species in having a more parallel-sided and narrower (tr.) glabella.

Description: — Glabella only slightly more than 1.6 times as long as wide, with subparallel sides, and subtrapezoidal in front, but not as markedly as the front of the glabella of *P. helgoeyensis* n.sp. Three pairs of lateral glabellar furrows. S1 geniculate, oblique backwards, and reaching the occipital furrow. S2 and S3 convex, S3 more convex than S2, oblique backwards, deepest near the axial furrow, and reaching about one-fourth across the glabella. Glabella covered with relatively small nodes. Occipital furrow distinct, simple and straight. Occipital ring as wide as glabella. Axial furrow well defined.

Other parts of the trilobite unknown.

Dimensions: — The holotype is 13 mm long and 7.8 mm wide (occipital ring not included).

Occurrence: — Probably the Upper Chasmops Shale (4b*y*), OSLO—ASKER (Oslo).

Affinities: — *P. perlongus* (BRØGGER) is probably closely related to both *P. exsul* (BEYRICH, 1846) and *P. gladiator* (EICHWALD, 1857), and may morphologically occupy an intermediate position between these. *P. perlongus* differs from *P. exsul* in having narrower glabella, and it differs from *P. gladiator* in having the glabella possessed with relatively large and conspicuous nodes.

SUBFAMILY CYRTOMETOPINAE ÖPIK, 1937

Remarks: — With the inclusion of *Cyrtometopella* n. gen. in the subfamily the diagonsis as given in the Treatise on Invertebrate Paleontology, Volume 0 may be changed to state that eye ridges are present or not (if present partly coalesced with the sutural ridges), and that the pygidium has 2 to 4 (instead of 3 to 4) pairs of pleural spines, the posterior pair being more or less reduced.

Genus *Cyrtometopella* n. gen.

Name: — Referring to the great likeness to the genus *Cyrtometopus*, I prefer to call this genus *Cyrtometopella*.

Type species: — *Cyrtometopella tumula* n. gen., n. sp.

Diagnosis: — A genus of the subfamily Cyrtometopinae with the cranidium subsemicircular in outline, glabella oval in outline, with three pairs of lateral glabellar furrows, reaching or nearly reaching the occipital furrow. Eyes located opposite or slightly behind S2. Eye ridges present or not (if present partly coalesced with sutural ridges).

Hypostome and thorax unknown.

Pygidium with two pairs of pleural spines, posterior pair reduced.

Affinities: — Cranidium resembling that of *Cyrtometopus*, but with the eyes opposite or slightly behind S2. The postocular fixigenae are narrower (tr.) and more convex transversely than those of *Cyrtometopus*. Glabella resembles that of *Cyrtometopus* in having convex sides, but the convexity is stronger both in the longitudinal and the transversal direction. Sutural ridges, if present, differs from those of *Cyrtometopus* in being partly coalesced with the eye ridges.

The pygidium differs distinctly from that of *Cyrtometopus* in having only two pairs of pleural spines.

The cranidium of *Cyrtometopella* thus resembles that of *Cyrtometopus* so much that it can hardly be doubted that the two genera are closely related. For this reason I place the genus *Cyrtometopella* in the subfamily Cyrtometopinae.

Remarks: — The species *Hemisphaerocoryphe? rosenthali* (Schmidt, 1881) shows a very close relationship to the two species *Cyrtometopella tumula* n. sp. and *C. askerensis* n. sp.. For this reason I propose to include *H.? rosenthali* in the genus *Cyrtometopella*.

Cyrtometopella tumula n. gen., n. sp.

Pl. 1, figs. 7—10.

Name: — From Latin *tumulus* (barrow, hillock), referring to the shape of the glabella.

Type data: — Holotype (here selected) is a cranidium (P.M.O. no. 72146) from the transitional beds between the Ampyx Limestone (4*aβ*) and the Lower Chasmops Shale (4b*a*) at Ildjernet in the Oslo—Asker district.

Material: — In addition to the holotype 4 more or less fragmentary cranidia (P.M.O. nos. 72147 (2 specimens), 72148, and 72149), and 2 pygidia (P.M.O. nos. 18910 and 72149).

Diagnosis: — A *Cyrtometopella* species with the cranidium about 3 times as wide as long, rather convex transversely. Glabella bulbous, and with three pairs of lateral glabellar furrows; S1 oblique backwards, S2 transverse, S3 oblique forwards. Occipital ring as wide as posterior end of glabella. Fixigenae about twice as wide as occipital ring, and with long genal spines. Eyes located slightly behind S2.

Librigenae, hypostome and thorax unknown.

Pygidium with 2 pairs of pleural spines, anterior pair strong, posterior pair reduced.

Description: — Cranidium subtriangular in outline, strongly convex transversely, only slightly more than 3 times as wide as long. Glabella oval in outline and very bulbous. Three pairs of lateral glabellar furrows. S1 convex, oblique backwards, reaching one-third across the glabella, and nearly reaching the occipital furrow. S2 almost straight, transverse, reaching only one-eight across the glabella. S3 straight, oblique forwards, less distinct. Glabella with numerous small nodes. Occipital furrow simple, straight, deep and distinct. Occipital ring about as wide as posterior part of glabella, covered with small and less distinct nodes. Mesial tubercle not developed. Preoccipital glabellar lobes suboval in outline. Axial furrow deep and well defined. Postocular fixigenae about twice as wide as occipital ring, with relatively long genal spines, and covered with nodes. The genal spines are round and slightly curved, oblique backwards. Posterior border furrow narrow (long.) near the axial furrow, but widening laterally. Posterior border narrow (long.). Width of interocular fixigenae about one-fourth of that of glabella at eye-line. Eyes located at transverse mid-line of the cranidium. Palpebral lobes convex and relatively small. Eye ridges not developed on present material. Preglabellar field narrow (long.). Preocular part of facial suture converging forwards, postocular part transverse.

Librigenae, hypostome and thorax unknown.

Pygidium about twice as wide as long, axis occupying about one-third of total width. Two pairs of pleural spines, anterior pair about 3 times as long as the pygidium proper. Posterior pair of pleural spines much shorter, only about one-fourth as long as the pygidium proper.

Dimensions: — The holotype is 7.1 mm long and 22 mm wide. The largest pygidium is 8 mm long and 15.5 mm wide (spines not included).

Occurrence: — Ampyx Limestone (4aβ) and Lower Chasmops Shale (4bα). OSLO—ASKER: Ildjernet (4aβ), Slependen (4bα), Vassholmen (4aβ).

Affinities: — *C. tumula* n. sp. is closely related to *C. rosenthali* (SCHMIDT, 1881) (here transferred to this genus), but differs from it in having a wider cranidium and a more bulbous glabella.

Cyrtometopella askerensis n. gen., n. sp.

Pl. 1, figs. 5—6.

Name: — From the district of Asker.

Type data: — Holotype (here selected) is a cephalon (P.M.O. no. 72226) from the Ampyx Limestone (4aβ) at Terneskjær, Asker.

Material: — In addition to the holotype only one fragmentary cranidium (P.M.O. no. 72224).

Diagnosis: — A *Cyrtometopella* species with cephalon subsemicircular in outline. Cranidium more than twice as wide as long. Glabella with convex sides and 3 pairs of lateral glabellar furrows. Fixigenae pitted, and provided with relatively long genal spines. Eyes located at transverse mid-line of the cranidium.

Other parts of the trilobite unknown.

C. askerensis n. sp. differs from the type species in having a well rounded convexity of the glabella and narrower and more convex fixigenae.

Description: — Cephalon subsemicircular in outline, rather convex transversely. Cranidium more than twice as wide as long. Glabella oval in outline, well rounded in front, and covered with numerous small nodes. Three pairs of lateral glabellar furrows, all deepest near the axial furrow. S1 convex, oblique backwards, and nearly reaching the occipital furrow. S2 slightly convex, transverse, and reaching about one-fifth across the glabella. S3 almost straight, oblique slightly forwards, and reaching about one-fifth across the glabella. Preoccipital glabellar lobe oval in outline. Occipital furrow broad (long.), deepest at the axial furrow, simple and straight. Occipital ring about as wide as posterior part of glabella, and covered with numerous small nodes. Mesial tubercle not developed. Axial furrow deep and distinct. Fixigenae about 1.5 times as wide as occipital ring, pitted only faintly near the glabella, but more strongly at the sides, and provided with relatively long genal spines. The genal spines are flattened and slightly curved, pointing obliquely backwards. Posterior border furrow deep and broad (long). Posterior border narrow (long.), but somewhat wider near the postero-lateral corner. Width of interocular fixigenae about one-third of that of glabella at eye-line. Eyes seem to be stalked, located at the transverse mid-line of the cranidium. Eye ridges partly coalesced with the sutural ridges, pointing obliquely backwards. Preglabellar field appears to be only a narrow (long.) brim. Preocular part of facial suture converging forwards, postocular part transverse.

Librigenae small, and with well rounded outer margin. Lateral border furrow deep and broad (tr.). Lateral border smooth and well defined.

Other parts of the trilobite are unknown.

Dimensions: — The holotype, which is the largest specimen present, is 7.7 mm long and 18 mm wide.

Occurrence: — Ampyx Limestone (4aβ). OSLO—ASKER: Terneskjær.

Affinities: — *C. askerensis* n. sp. is closely related to both *C. tumula* n. sp. and *C. rosenthali* (SCHMIDT, 1881) (here transferred to this genus). The cranidium differs from that of *C. tumula* in being narrower (tr.) and in having a better rounded glabella in lateral view. It differs from that of *C. rosenthali* in having the eyes further forwards and closer to the glabella.

SUBFAMILY ACANTHOPARYPHINAE WHITTINGTON & EVITT, 1953

Genus *Nieszkowskia* SCHMIDT, 1881

Type species: — *Sphaerexochus cephaloceros* NIESZKOWSKI, 1857

Nieszkowskia norvegica n. sp.

Pl. 2, figs. 1–2.

Name: — From Latin *norvegicus* (from Norway).

Type data: — Holotype (here selected) is a fragmentary cranidium (P.M.O. no. 8675) from the Echinosphaerites Beds (4b*a*?) at Skjellbukta, Frierfjord, Skien—Langesund.

Material: — In addition to the holotype only one fragmentary cranidium (P.M.O. no. 20333).

Diagnosis: — A *Nieszkowskia* species with a broad and strong glabella, and strong glabellar spine. The whole cranidium is covered with both larger and smaller nodes. Glabella tapering forwards.

No other parts of the trilobite are known.

N. norvegica n. sp. differs from the type species in having a wider glabella and a stouter glabellar spine.

Description: — Glabella about 1.4 times as long as wide, with straight sides converging forwards, and bluntly rounded in front. Three pairs of lateral glabellar furrows. S1 strongly convex, oblique backwards, and not reaching the occipital furrow. S2 and S3 convex, oblique backwards, and reaching about one-fourth across the glabella. Between the preoccipital glabellar lobes there is a long and strong spine, gently curved and pointing backwards. Occipital furrow simple, straight and well developed. Occipital ring only slightly narrower than posterior part of the glabella, broad (long.) at the axial line, but tapering to the sides. Mesial tubercle less distinct. Axial furrow distinct, but not very deep. Fixigenae not completely preserved on present material. Posterior border furrow narrow (long.), deep and distinct. Posterior border broad (long.). Preglabellar field narrow (long.). The whole cranidium (glabellar spine included) densely covered with both small and large nodes.

Other parts of the trilobite are unknown.

Dimensions: — The glabella of the holotype, which is the largest specimen present, is 34 mm long and 24.2 mm wide.

Occurrence: — Ampyx Limestone (4aβ) and Echinosphaerites Beds (4b*a*?). OSLO—ASKER: Søndre Kojatangen (4aβ), SKIEN—LANGESUND: Skjellbukta (Ech.).

Affinities: — *N. norvegica* n. sp. is closely related to *N. variolaris* (LINNARSSON, 1869), but differs from it in having the glabella less tapering forwards, more bluntly rounded in front, and in having a longer and stronger glabellar spine.

Furthermore it seems to be related to *N. cephaloceros* (NIESZKOWSKI, 1857), but differs from it in the characters mentioned in the diagnosis.

SUBFAMILY DEIPHONINAE RAYMOND, 1913

Genus *Hemisphaerocoryphe* REED, 1896

Type species: — *Sphaerexochus pseudohemicranium* NIESZKOWSKI, 1859.

Hemisphaerocoryphe inflatus n.sp.

Pl. 1, figs. 11–12.

Name: — From Latin *inflatus* (puffed up, swollen).

Type data: — Holotype (here selected) is a small but well preserved cranidium (P.M.O. no. 67357) from the Coelosphaeridium Beds (4bβ?) at Eina, Toten.

Material: — In addition to the holotype 5 more or less fragmentary cranidia (P.M.O. nos. 5352, 63386, 72222, 72223, and 72254).

Diagnosis: — A *Hemisphaerocoryphe* species with a cranidium more than twice as wide as long. Glabella bulbous, nearly spherical, 3 pairs of lateral glabellar furrows, all oblique backwards. Fixigenae relatively wide, probably with genal spines. Eyes located at transverse mid-line of the cranidium.

Other parts of the trilobite are not known.

H. inflatus differs from the type species in having a more bulbous glabella, narrower occipital ring, and the eyes located further backwards.

Description: — Cranidium only slightly less than 2.4 times as wide as long. Glabella very bulbous, nearly spherical, leaning somewhat forward, and faintly granulated. Three pairs of lateral glabellar furrows. S1 convex, oblique backwards, deep, distinct, and reaching the occipital furrow. S2 convex, oblique backwards, short, deepest a short distance from the axial furrow. S3 pit-like, less distinct. Preoccipital glabellar lobe subellipsoidal in outline. Occipital furrow deep and distinct, simple and convex, deepest near the axial furrow. Occipital ring narrow (long.), width about that of glabella between the preoccipital glabellar lobes, and probably granulated. Mesial tubercle not developed. Axial furrow deep and distinct. Postocular fixigenae slightly wider (tr.) than the occipital ring, pitted less distinct near the axial furrow, but stronger laterally, and probably provided with genal spines. Posterior border furrow broad (long.), deep and well defined. Posterior border narrow (long.) and distinct. Eyes located at transverse mid-line of the cranidium. Palpebral lobes medium-sized, convex and protuberance-like. Width of interocular fixigenae about one-fourth of that of glabella at eye-line. Eye ridges straight and distinct, oblique backwards, narrow (long.) near the glabella, but widening laterally. Preglabellar field reduced to be only a narrow (long.) brim on glabella. Preocular part of facial suture converging forwards, postocular part transverse.

Other parts of the trilobite are unknown.

Dimensions: — The holotype is 6.5 mm long and 15.5 mm wide. The largest fragmentary cranidium is about twice as large as the holotype.

Occurrence: — Lower Chasmops Shale (4b*α*) and Coelosphaeridium Beds (4b*β*?). TOTEN: Eina (Coel.), OSLO—ASKER: Gåsøya (4b*α*).

Affinities: — *H. inflatus* n. sp. is closely related to *H. granulatus* (ANGELIN, 1854), and may possibly be a subspecies of this. It differs from *H. granulatus* in having broader (long.) fixigenae.

Hemisphaerocoryphe sphaericus (ESMARK, 1833)

Pl. 2, fig. 3.

1833 *Trilobites sphaericus* n. sp. — ESMARK, p. 269, pl. VII, fig. 8. (Descr. and figs. of cranidium).

1940 *Hemisphaerocoryphe sphaericus* (ESMARK) — STØRMER, p. 124, text figs. 2—3, pl. 1, figs. 6—7. (Remarks and figs. of cranidium).

Type data: — Neotype is the cranidium (P.M.O. no. 61747) figured by Størmer (1940), from the Ampyx Limestone (4a*β*) or the Lower Chasmops Shale (4b*α*) at Løken, Bygdøy in Oslo.

Material: — In addition to the neotype 4 incomplete cranidia are preserved at Paleontological Museum in Oslo (P.M.O. nos. 3688, 18862, 20425, and 56174).

Diagnosis: — A *Hemisphaerocoryphe* species with cephalon subsemicircular in outline, cranidium more than twice as wide as long. Three pairs of lateral glabellar furrows, S1 and S2 oblique backwards, S3 oblique forwards. Fixigenae as wide as occipital ring, pitted. Eyes located slightly in front of the transverse mid-line of the cranidium.

Other parts of the trilobite are unknown.

H. sphaericus (ESMARK) differs from the type species in having a more bulbous glabella, well defined preoccipital glabellar lobes, and the eyes farther from the glabella.

Description: — Cephalon subsemicircular in outline, strongly convex transversely. Cranidium triangular in outline, and about 2.2 times as wide as long. Glabella strongly convex, about 1.5 times as long as high, well rounded in front, and covered with relatively small nodes all over. Three pairs of lateral glabellar furrows. S1 deep and distinct, convex, oblique backwards, and reaching the occipital furrow. S2 deep, but not as deep as S1, convex, oblique backwards, reaching one-sixth across the glabella. S3 slightly convex, nearly straighth, oblique forwards, less distinct. Preoccipital glabellar lobe suboval in outline. Occipital furrow deep and distinct, simple and convex. Occipital ring narrow (long.), and as wide as glabella between the preoccipital glabellar lobes, probably granulated. Mesial tubercle not developed. Axial furrow deep and well defined. Postocular fixigenae pitted, and as wide as occipital ring, genal spines appear to be present. Posterior border furrow deep and broad (long.). Posterior border narrow (long.). Eyes located slightly in front of the transverse mid-line of the cranidium. Palpebral lobes medium-sized, well defined. Eye ridges not present, or probably coalesced with the sutural ridges. Width of interocular fixigenae about one-third of that of glabella at eye-line. Preglabellar field reduced to a narrow brim on the glabella. Preocular part of facial suture converging forwards, postocular part transverse.

Other parts of the trilobite unknown.

Dimensions: — The neotype, which is the largest specimen present, is 8.9 mm long and 22.3 mm wide (occipital ring and genal spines missing).

Occurrence: — Ampyx Limestone (4aβ) and Lower Chasmops Shale (4b*a*). Oslo—Asker: Royal Castle, Bygdøy, Bjerkøy, Ostøya.

Affinities: — *H. sphaericus* (ESMARK) is closely related to *H. pseudohemicranium* (NIESZKOWSKI, 1859), but differs from it in the characters mentioned in the diagnosis.

Family Pliomeridae RAYMOND, 1913

[*nom. transl.* ÖPIK, 1937 (*ex* Pliomerinae RAYMOND, 1913)]
[= Amphionidae PICTET, 1854]

SUBFAMILY PLIOMERINAE RAYMOND, 1913

Genus *Pliomerops* RAYMOND, 1905

Type species: — *Amphion canadensis* BILLINGS, 1859

Pliomerops sp.

Pl. 2, fig. 4.

Material: — Only one fragmentary cranidium (P.M.O. no. 36600) and counterpiece (P.M.O. no. 36599) from the Ogygiocaris Shale (4a*a*) at Redalen in Biri (outside the Oslo region).

Description: — Cranidium about 2.5 times as wide as long. Glabella wide in front, but tapering backwards. S1 slightly convex, oblique slightly backwards. S2 and S3 straight, transverse, deepest nearest to the axial line, and reaching more than one-third across the glabella. Preoccipital glabellar lobes small and narrow (long.). Occipital ring poorly preserved, but seems to show a strong mesial tubercle. Postocular fixigenae wide (tr.) and broad (long.), and strongly reticulate pitted. Posterior border furrow deep and distinct. Posterior border well defined. Eyes seem to be located far forwards, opposite S3. Postocular part of facial suture gently curved, directed transverse.

Other parts of the trilobite are unknown.

Dimensions: — The only known cranidium is 12 mm long and 30.4 mm wide.

Occurrence: — Ogygiocaris Shale (4a*a*). Redalen in Biri (North of the Oslo region).

Family Encrinuridae ANGELIN, 1854

SUBFAMILY ENCRINURINAE ANGELIN, 1854

[*nom. transl.* HUPÉ, 1955 (*ex* Encrinuridae ANGELIN, 1854)]

Genus *Encrinurus* EMMRICH, 1844

Type species: — *Entomostracites punctatus* WAHLENBERG, 1821

Encrinurus sp.

Pl. 2, fig. 5.

Material: — Only one fragmentary pygidium (P.M.O. no. 72150) from the Mjøsa Limestone ($4b\delta_2$) at Bergevika, Helgøya.

Description: — Pygidium about 1.3 times as long as wide, with at least 10 pairs of pleural lobes. Axis occupying about one-third of total width. The pleural lobes are provided with short spines. Pleural region strongly convex. Posterior spine lacking. Axis with at least 24 small axial rings.

Other parts of the trilobite unknown.

Dimensions: — The pygidium is 6 mm long and 4.5 mm wide.

Occurrence: — Mjøsa Limestone ($4b\delta_2$). NES—HAMAR: Helgøya.

Affinities: — *E.* sp. seems to be closely related to *E. ? moë* (MÄNNIL, 1958), but differs from it in having narrower pygidium.

Remarks: — *E.* sp. is undoubtedly a new species, but there is no reason to name it until better material is available.

SUBFAMILY CYBELINAE HOLLIDAY, 1942

[nom. transl. HUPÉ, 1955 (by error as "sous-famille de Cybelidae") (*ex* Cybelidae HOLLIDAY, 1942)] [*emend.* HUPÉ, 1955]

Genus *Atractopyge* HAWLE & CORDA, 1847

Type species: — *Calymene verrucosa* DALMAN, 1827

Atractopyge brevicauda (ANGELIN, 1854)

Pl. 2, fig. 6—7.

1854 *Cybele brevicauda* n. sp. — ANGELIN, p. 89, pl. XLI, fig. 14. (Short diagn. and fig. of pygidium).
?1857 *Zethus brevicauda* n. sp. — NIESZKOWSKI, p. 617.
?1857 *Zethus bellatulus* n. sp. — NIESZKOWSKI, p. 613.
?1857 *Zethus atractopyge* n. sp. — NIESZKOWSKI, p. 616.
?1857 *Encrinurus sexcostatus* n. sp. — NIESZKOWSKI, p. 610.

1881 *Cybele brevicauda* ANGELIN? — SCHMIDT, p. 219, pl. XI, fig. 23, pl. XIV, figs. 7—10, pl. XV, figs. 15—17. (Descr. and figs. of cranidia and pygidia).
1884 *Cybele brevicauda* ANGELIN — TÖRNQUIST, p. 26. (Recorded).
1897 *Cybele brevicauda* ANGELIN — KIÆR, p. 13. (Recorded).
1909 *Cybele brevicauda* ANGELIN — HOLTEDAHL, pp. 20, 31, 32, 40. (Recorded).
1925 *Cybele brevicauda* ANGELIN — WARBURG, p. 337, pl. XI, fig. 38. (Remarks and fig. of holotype).

Type data: — As holotype (by monotypy) I select the pygidium described by ANGELIN (1854, pl. XLI, fig. 14) from Osmundsberg, Sweden, and preserved in Naturhistoriska Riksmuseet, Stockholm.

Norwegian material: — Only two fragmentary cranidia (P.M.O. nos. 67922 and 72255) and two incomplete pygidia P.(M.O. nos. 7384 and 35426) are present in the Middle Ordovician, but cranidia, pygidia and thoracic shields are also known from the Upper Ordovician of the Oslo region.

Diagnosis (*based on Norwegian material*): — An *Atractopyge* species with the glabella strongly expanding forwards, concave sides, and covered with large nodes. Three pairs of lateral glabellar furrows, all oblique backwards. Fixigenae rather convex. Eyes located in posterior part of the cranidium.

Pygidium about as long as wide. Axis strongly arched down in lateral view.

A. brevicauda (ANGELIN) differs from the type species in having larger eyes, located in the posterior half part of the cranidium, glabella stronger convex, and narrower (long.) fixigenae. The pygidium differs in having small pleural spines, and a smoother axis.

Description of Norwegian material: — The description is based on the Norwegian Middle Ordovician material.

Cranidium subtriangular in outline. Glabella with slightly concave sides, strongly expanding forwards, and well rounded and provided with a row of relatively large and well defined tubercles in front. Three pairs of lateral glabellar furrows. S1 convex, oblique backwards, nearly reaching the occipital furrow, and deepest a short distance from the axial furrow. S2 and S3 slightly convex, oblique backwards, deepest a short distance from the axial furrow, and reaching one-third across the glabella. Preoccipital glabellar lobe small and less distinct. Occipital furrow faint and poorly defined. Occipital ring as wide as posterior part of glabella. Mesial tubercle not preserved on present material. Postocular fixigenae strongly convex, width about that of occipital ring. Eyes located in posterior half of the cranidium. Eye ridges not developed. Width of interocular fixigenae about one-half of that of glabella at eye-line. A deep pit-like impression is observed posterior to the antero-lateral corner of the glabella. The whole cranidium is covered with large nodes.

Librigenae, hypostome and thorax not present in the Middle Ordovician material.

Pygidium about as wide as long. Pleural region strongly convex. The ribs are provided with short spines, which are broad-based, but taper rapidly. Axis occupying about one-third of total width, strongly arched down in lateral view, and with 5 more or less distinct nodes.

Dimensions: — The cranidium is 10.7 mm long and 14.2 mm wide. The largest pygidium is 11.8 mm wide and 11.2 mm long.

Occurrence in Norway: — Upper Chasmops Limestone (4bδ), Gagnum Shale (4bδ?), and Mjøsa Limestone (4bδ). NES—HAMAR: Helgøya (Mjøsa Lmst.), HADELAND: Sølvsberget (4bδ), Grimsrud (Gagnum Sh.), RINGERIKE: Røysetangen (4bδ).

Affinities: — *A. brevicauda* (ANGELIN) is closely related to *A. kutorgae* (SCHMIDT, 1881), but differs from it in having the glabella less expanding forwards.

The pygidium differs in being narrower (tr.).

Atractopyge dentata (ESMARK, 1833)

Pl. 3, figs. 1—2, pl. 4, figs. 1—4.

1833 *Trilobites dentatus* n. sp. — ESMARK, p. 269, pl. VII, fig. 13. (Descr. and fig. of pygidium).

1854 *Cybele dentata* ESMARK — ANGELIN, p. 89, pl. XLI, fig. 12. (Short diagn. and fig. of pygidium).

1881 *Cybele Grewingki* n. sp. — SCHMIDT, p. 211, text fig. 11, pl. XIV, figs. 1—2. (Descr. and figs. of cranidium and thorax. Non fig. of pygidium).

1887 *Cybele* sp. — BRØGGER, p. 19. (Recorded).

1888 *Cybele Grewingki* SCHMIDT — WIGAND, p. 89, pl. X, fig. 17. (Descr. and fig. of cranidium).

1907 *Cybele Grewingki* SCHMIDT — SCHHMIDT, p. 20. (Mentioned).

1909 *Cybele Grewingki* SCHMIDT — HOLTEDAHL, pp. 16, 30, 40. (Recorded).

1937 *Cybele* (*Cybelella*) *grewingki* SCHMIDT — ÖPIK, p. 123, text fig. 36. (Discussion and fig. of "holotype").

1940 *Cybele dentata* (ESMARK) — STØRMER, p. 125, text fig. 2, pl. 1, fig. 5. (Remarks and fig. of lectotype).

1940 *Cybele dentata* (ESMARK) — GRORUD, p. 160, pl. 2, fig. a. (Recorded and fig. of pygidium).
1953 *Cybele* sp. — STØRMER, pp. 62, 64, 91. (Recorded).

Type data: — Lectotype is the pygidium (P.M.O. no. 56161), figured by STØRMER (1940, pl. 1, fig. 5) from the Ogygiocaris Shale (4a*a*) or the Ampyx Limestone (4a*a*) at the Royal Palace in Oslo.

Norwegian material: — Two nearly complete specimens (P.M.O. nos. 61609 and 72151), one specimen with cranidium and thorax preserved (P.M.O. no. 18831 and counterpiece no. 18832), 13 cranidia (P.M.O. nos. 1940, 1946, 1980, 2120, 18833, 20366, 20626, 56509, 63187, 63253, 63255, 72184, 72204), 6 librigenae (P.M.O. nos. 7383, 34610, 34611, 5277, 55975, 72211), and 20 pygidia (P.M.O. nos. 6346, 6347, 6898, 7377, 7382, 7519, 20625, 33473, 36979, 56042, 56065, 56089, 56161 a, 56259, 56315, 56440, 56508, 62111, 72162, 72170) are preserved at the Paleontological Museum in Oslo.

Diagnosis: — An *Atractopyge* species with cephalon subsemicircular in outline, rather convex transversely. Cranidium about 2.5 times as wide as long. Glabella with subparallel sides, only slightly wider and with 5 distinct tubercles or short spines in front, 4 pairs of more or less distinct nodes are observed along the axial line. Three pairs of lateral glabellar furrows, all oblique backwards. Occipital ring as wide as posterior part of glabella. Fixigenae about twice as wide as occipital ring. Genal spines lacking. Eyes stalked, located slightly behind the transverse mid-line of the cranidium.

Librigenae pitted, lateral margin well rounded.

Thorax with 12 segments. Pleurae with long spines. The spines on the 6. segment are as long as the total length of the dorsal shield, and parallel to the axis.

Pygidium with 4 pairs of ribs, provided with short spines. Axis occupying about one-third of total width, and only slightly arched down in lateral view.

A. dentata (ESMARK) differs from the type species in having narrower (tr.) fixigenae, wider (tr.) librigenae, and stalked eyes, located further backwards and closer to the glabella.

The pygidium differs in being narrower, and in having small pleural spines.

Description: — The description is based on the Norwegian material.

Dorsal shield suboval in outline, somewhat narrower (tr.) in the pygidal region. Cephalon subsemicircular in outline, rather convex transversely. Cranidium about 2.5 times as wide as long. Glabella with subparallel sides, only slightly wider in front, and bearing 8 more or less distinct nodes located in 2 longitudinal lines. Five distinct tubercles or short spines in front of glabella. The spines are gently flattened and broad-based, but taper rapidly. Three pairs of lateral glabellar furrows. S1 geniculate, oblique backwards, deep and short, and not reaching the occipital furrow or the the axial furrow. S2 pit-like, nearly oval, oblique backwards. S3 straight, more oblique backwards than S2, not reaching the axial furrow. Occipital furrow well defined, simple and convex. Occipital ring as wide as posterior part of the glabella, with a distinct mesial tubercle. Furthermore 2 small more or less distinct nodes on each side of the mesial tubercle. Axial furrow faint. Postocular fixigenae pitted, faint near the glabella, but stronger laterally, width only slightly less than twice the width of the occipital ring. Behind the eyes some more or less distinct nodes are to be observed. Posterior border furrow relatively faint and narrow (long.) near the axial furrow, but widening and deep laterally. Posterior border narrow (long.) and well defined. Genal spines lacking. Eyes stalked (maximum length 8 mm), located slightly behind the transverse mid-line of the cranidium. Eye ridges straight and well defined, directed obliquely backwards. Width of interocular fixigenae about one-half of that of glabella at eye-line. Anterior pits deep, slightly elongate, situated ahead of the most anterior lateral glabellar furrows. One large node is located outside the antero-lateral corner of the glabella. Preglabeller field narrow (long.). Preocular part of facial suture converging forwards, postocular part transverse.

Librigenae reticulate pitted, strongly arched, provided with some distinct nodes located near to the eyes. Lateral margin well rounded. Lateral border broad (tr.) and with a row of large nodes. Lateral border furrows faint.

Hypostome unknown.

Thorax with 12 segments. Axis strongly convex transversely. Pleurae bearing transverse furrows and long spines. The spines on the 1.—5. segments are directed straight outwards or slightly oblique backwards. The spines on the 6. segment are nearly parallel to the axis, gently curved, and about as long as the total length of the dorsal shield. The spines on the 7.—12. segments are parallel to the spines on the 6. segment, and about one-fourth as long as these.

Pygidium with 4 pairs of ribs, which are provided with short spines. The spines are round and broad-based, but taper rapidly. Axial ring not well defined. Postero-lateral corners shoulder-like, lateral margin slightly convex, and converging backwards. Axis occupying about one-third of total width, only slightly arched down in lateral view, and with at least 14 pairs of small furrows. Ribs with few more or less distinct nodes.

Dimensions: — The largest complete specimen is 62 mm long and 38 mm wide (spines on the 6. segment not included). The largest cranidium has an axial length of 23 mm and a width of 61 mm.

The largest pygidium, which belongs to the largest complete specimen, is 16 mm long and 15 mm wide.

Occurrence: — Ogygiocaris Shale (4a*α*), Ampyx Limestone (4a*β*), Cephalopod Shale (4a*β*?), Lower Chasmops Shale (4b*α*), Echinospaerites Beds (4b*α*?), and Lower Chasmops Limestone (4b*β*). NES—HAMAR: Hovindsholm, HADELAND: Gran, RINGERIKE Bratterud, Ringsås, Røysetangen, OSLO—ASKER: Bygdøy, Sinsen, Skarpsno, Slottsbakken, Smestad, Tørtberg, Billingstad, Delebukta, Nakkholmen, Nesøya, Ostøya, Slependen, LANGESUND—SKIEN: Frierfjorden.

Affinities: — *A. dentata* (ESMARK) is closely related to *A. coronata* (SCHMIDT, 1881), but differs from it in having shorter spines in front of the glabella, narrower fixigenae, and the eyes located in the posterior half of the cranidium.

The pygidium differs in being narrower and in having shorter spines.

Remarks: — I believe that the species *A. grewingki* (SCHMIDT, 1881) is a synonym of *A. dentata*. The holotype (selected by ÖPIK) described by SCHMIDT as *Cybele Grewingki* n. sp. is a specimen with cephalon and thorax preserved. This specimen shows the spines on the 6. segment that are so typical of *A. dentata*. The diagnosis and description he gives for the cephalon and thorax agrees well with the diagnosis and description of *A. dentata*. He also mentions that *C. Grewingki* is larger than any others of the Baltic *Cybele* species. The dimensions he gives agree well with the Norwegian material of *A. dentata*. The pygidium, which he describes as *C. Grewingki*, seems to belong to *A. kurtorgae* (SCHMIDT, 1881).

A. dentata also resembles the species *A. revaliensis* (SCHMIDT, 1881) so much that I cannot find any differing characters between them. Therefore I believe that *A. revaliensis* is a synonym of *A. dentata.*

Atractopyge gracilis n.sp.

Pl. 4, figs. 5–9.

1887 *Cybele* sp. — BRØGGER, p. 23. (Recorded).

Name: — From Latin *gracilis*, (slender, thin).

Type data: — Holotype (here selected) is a cranidium (P.M.O. no. 6531) from the Upper Chasmops Limestone (4bδ) at Terneholmen, Asker.

Material: — In addition to the holotype 9 cranidia (P.M.O. nos. 5600, 5615, 5917, 5937, 5944, 6523, 20443, 65146, 72192), 2 librigenae (P.M.O. nos. 5818, 72145), and 11 pygidia (P.M.O. nos. H 227, 5575, 5638, 5818, 5926 (2 specimens), 5927, 9095, 72190 (2 specimens), 72191), are preserved at the Paleontological Museum in Oslo.

Diagnosis: — An *Atractopyge* species with the cranidium about 3 times as wide as long, subparallel-sided glabella, and 3 pairs of lateral glabellar furrows, all oblique backwards. Occipital ring as wide as glabella. Fixigenae relatively wide, strongly granulated. Eyes stalked, located slightly behind the transverse mid-line of the cranidium. Genal spines relatively long, broad (tr.) and flattened.

Librigenae granulated.

Hypostome and thorax are unknown.

Pygidium slightly longer than wide, axis occupying about one-third of total width, and strongly arched down in lateral view.

A. gracilis n. sp. differs from the type species in having a subparallel-sided glabella, long genal spines, and the eyes further backwards and closer to the glabella.

The pygidium differs in being narrower, and in having small pleural spines.

Description: — Cranidium subtriangular in outline, about 3 times as wide as long, rather convex transversely. Glabella with subparallel sides, strongly granulated. Three pairs of lateral glabellar furrows, all reaching about one-third across the glabella. S1 geniculate, oblique backwards, deepest a short distance from the axial furrow, and reaching the occipital furrow. S2 elongate, nearly pit-like, oblique backwards, and not reaching the axial furrow. S3 almost straight, more oblique backwards than S2, and deepest a short distance from the axial furrow. Preoccipital glabellar lobes elongate in outline. Occipital furrow composite, deep and well defined, and with a deep and pit-like impression a short distance from the axial furrow. Occipital ring as wide as glabella, and with a well developed and keel-like mesial tubercle. Furthermore some small nodes on each side of the mesial tubercle. Axial furrow deep and broad (tr.). Postocular fixigenae about twice as wide as occipital ring, and strongly granulated. Genal spines relatively long, flattened and broad-based. Inner half of posterior border narrow (long.), outer half broad. Posterior border furrow deep and narrow (long.) near the axial furrow, but broader and fainter laterally. Eyes stalked, located slightly behind the transverse mid-line of the cranidium. Eye ridges well developed, straight, directed obliquely backwards. Width of interocular fixigenae about one-half of that of glabella at eye-line. Anterior pits deep, situated ahead of the most anterior lateral glabellar furrows. Preglabellar field possibly with long spines in front because the anterior margin is broken on all specimens. Preocular part of facial suture converging forwards, postocular part running convex transversely.

Librigenae faintly granulated and strongly arched. Lateral margin well rounded. Lateral border relatively broad (tr.) and smooth. Lateral border furrow broad (tr.) and deep.

Hypostome and thorax unknown.

Pygidium only slightly less than 1.3 times as long as wide, straight in front, and with the lateral margins convex, converging backwards. Four pairs of ribs provided with short and broad-based round spines. Axis occupying about one-third of the total width, strongly arched down in lateral view, and with a large number of small transverse furrows. Both axis and pleurae with few small nodes.

Dimensions: — The holotype, which is the largest specimen present, is 27.6 mm wide and 9.1 mm long. The genal spines are 12 mm long.

The largest pygidium is 7.5 mm long and 6.0 mm wide.

Occurrence: — Upper Chasmops Limestone (4dδ) and Sphaeronid Limestone (4bδ?). HADELAND: S. Gran (Sphaer.), OSLO—ASKER: Ostøya, Terneholmen (4bδ).

Affinities: — *A. gracilis* n. sp. is closely related to *A. rex* (NIEszkowski, 1857), but differs from it in having long genal spines and the eyes more posteriorly situated.

The pygidium differs in being narrower and more arched down in lateral view.

A. gracilis n. sp. also seems to be related to *A. coronata* (Schmidt, 1881), but differs from this in having the eyes further backwards and closer to the glabella, eye ridges directed obliquely backwards, and long genal spines.

The pygidium differs in being narrower.

Atractopyge kutorgae (Schmidt, 1881)

Pl. 2, figs. 8–9.

1881 *Cybele Kutorgae* n. sp. — Schmidt, p. 217, pl. XV, figs. 11–14, pl. XVI, figs. 39a-b. Text figs. 12 a-b. (Descr. and figs. of cranidia and pygidia).

1907 *Cybele Kutorgae* Schmidt — Schmidt, p. 21, pl. I, figs. 16–18. (Remarks and figs. of cranidia).

Type data: — As lectotype should be selected one of the specimens described and figured by Schmidt, 1881.

Norwegian material: — Two fragmentary cranidia (P.M.O. nos. 34241 and 72153 with counterpiece 72152) are preserved at the Paleontological Museum in Oslo.

Diagnosis (based on Norwegian material): — An *Atractopyge* species with cranidium slightly more than 2.5 times as wide as long, rather convex transversely, and strongly granulated. Glabella strongly expanding forwards, and with a row of tubercles in front. Three pairs of short lateral glabellar furrows. Occipital ring wider than posterior part of glabella. Fixigenae slightly wider than occipital ring. Genal spines probably present.

Other parts unknown in the Norwegian material.

A. kutorgae (Schmidt) differs from the type species in having the glabella more expanding forwards, and the eyes closer to the glabella.

Description of Norwegian material: — Cranidium subtriangular in outline, only slightly more than 2.5 times as wide as long, rather convex transversely, and covered with large nodes all over. The nodes proper carry smaller nodes. Glabella strongly expanding forwards, and with a row of tubercles in front. One large node outside the antero-lateral corners of the glabella. Three pairs of short lateral glabellar furrows, but they are longer and deeper on exfoliated cranidia. S1 slightly convex, oblique backwards, deepest at the axial furrow, and not reaching the occipital furrow. S2 almost straight, transverse, reaching one-fourth across the glabella. S3 stronger than S2, almost straight, oblique forwards, and deepest a short distance from the axial furrow. Occipital furrow well defined, simple and convex, and deepest at the axial furrow. Occipital ring slightly wider than posterior part of glabella. Mesial tubercle appears to be lacking. Axial furrow well defined. Postocular fixigenae only slightly wider than occipital ring. Posterior border furrow deep and narrow (long.) near the axial furrow, but wider and fainter laterally. Posterior border relatively broad (long.). Genal spines appear to be present. Eyes stalked, situated slightly behind the transverse mid-line of the cranidium. Eye ridges relatively distinct, straight, directed obliquely backwards. Width of interocular fixigenae about one-half of that of glabella at eye-line. Anterior pits deep, situated ahead of the most anterior glabellar furrows. Preglabellar field narrow (long.). Preocular part of facial suture converging forwards, postocular part transverse.

Dimensions: — The largest cranidium is 11 mm long and 28 mm wide.

Occurrence: — Lower Chasmops Shale (4b*a*) and Cyclocrinus Beds (4b*y*?). OSLO—ASKER: Bernhard Herres vei (4b*a*), HADELAND: Røykenvik (Cycl.).

Affinities: — As pointed out by SCHMIDT, 1881, *A. kutorgae* closely resembles *A. brevicauda* (ANGELIN, 1854), and seems to be a forerunner of this. This agrees well with the Norwegian material, but the two species have so few and so small differing characters that they may be only subspecies of the same species.

A. kutorgae also seems to be related to *A. woerthi* (EICHWALD, 1840), but differs from it in having a wider cranidium and the eyes located further backwards.

Atractopyge aff. *kutorgae* (SCHMIDT, 1881)

Pl. 2, figs. 10.

Material: — Only one large pygidium (P.M.O. no. 21961) is preserved at the Paleontological Museum in Oslo.

Affinities: — The pygidium closely resembles that of *A. kutorgae* (SCHMIDT, 1881), but differs in some characters. The pygidium of

A. kutorgae is slightly wider than long, maximum length (cf. SCHMIDT, 1881) 10.5 mm. The Norwegian pygidium is 26 mm long, and slightly longer than wide. It also differs in converging more rapidly backwards.

Occurrence: — Coelosphaeridium Beds (4bβ?). NES—HAMAR: Furuberget.

Atractopyge sp.

Material: — Only one fragmentary cranidium (P.M.O. no. 72252) and one librigena (P.M.O. no. 72251).

Description: — Cranidium more than twice as wide as long. Glabella with subparallel sides, probably with spines in front. Three pairs of deep and distinct lateral glabellar furrows. S1 slightly convex, oblique backwards. S2 almost straight, oblique backwards. S3 slightly convex, oblique backwards. S1, S2 and S3 deepest near the axial furrow. Occipital furrow deep and distinct, simple and straight. Occipital ring not present. Postocular fixigenae about twice as wide as glabella. Posterior border furrow broad (long.). Posterior border relatively narrow (long.). Eyes possibly stalked. Eye ridges not preserved. Width of interocular fixigenae about two-thirds of that of glabella at eye-line. Cranidium granulated all over.

Librigenae strongly granulated. Lateral margin well rounded.

Other parts of the trilobite are unknown.

Dimensions: — The cranidium is 4 mm long and 8.9 mm wide.

Occurrence: — Ampyx Limestone (4aβ). OSLO—ASKER: Bygdøy.

Affinities: — *A.* sp. is closely related to *A. rex* (NIESZKOWSKI, 1857), but differs from it in having a more parallel-sided glabella, and in having more and larger nodes all over the cranidium.

Subfamily Staurocephalinae PRANTL & PŘIBYL, 1947

Genus *Staurocephalus* BARRANDE, 1846

Type species: — *Staurocephalus murchisoni* BARRANDE, 1846

Staurocephalus sp.

1957 *Staurocephalus* sp. — KIELAN, p. 167, text fig. 4, pl. IV, fig. 3. (Descr. and fig. of cranidium).

Material: — Only one fragmentary cranidium (P.M.O. no. 67044).

Description and Dimensions: — See Kielan, 1957.

Occurrence: — Upper Chasmops Limestone (4bδ). OSLO—ASKER: Terneholmen.

REFERENCES

ANGELIN, N. P., 1854: Palæontologia scandinavica. Pars. I. Crustacea formationis transitionis. Fasc. II. — Pp. I—IX, 21—92, pls. XXV—XLI. Holmiæ (Stockholm).

BEYRICH, E., 1846: Untersuchungen über Trilobiten. Berlin. Pp. 1—37, pls. 1—4.

BRØGGER, W. C., 1882: Die Silurischen Etagen 2 und 3 im Kristianiagebiet und auf Eker. — Universitätsprogramm für 2 Sem. 1882, pp. I—VIII, 1—376, pls. I—XII, Kristiania (Oslo).

— 1887: Geologisk kart over øerne ved Kristiania — Nyt Mag. for Naturvid., XXXI, II, pp. 1—36. Kristiania (Oslo).

EICHWALD, E., 1840: Die Urwelt Russlands, durch Abbildungen erläutet. — St. Petersburg. Pp. 1—106, pls. 1—3.

ESMARK, H. M. T., 1833: Om nogle nye Arter af Trilobiter. — Mag. for Naturvid., 2. Række, vol. 1 (= whole series, vol. 11), pp. 268—270, pl. VII. — Christiania (Oslo).

GRORUD, H.-F., 1940: Et profil gjennem Ogygiaskifer og Ampyxkalk på Tørtberg, Frogner ved Oslo. — Norsk geol. tidsskr., 20, pp. 37—141, figs. 1—2. Oslo.

HENNINGSMOEN, G., 1957: The trilobite family Olenidae. — Skr. Norske Vidensk.-Akad., I kl., 1957, no. 1, pp. 1—303, pls. 1—31. Oslo.

HOLTEDAHL, O.: 1909: Studien über die Etage 4 des norwegischen Silursystems beim Mjøsen. — Skr. Norske Vidensk.-Akad., I kl., 1909, no. 7, pp. 1—76. Christiania (Oslo).

JAANUSSON, V., 1956: On the Trilobite Genus Celmus Angelin, 1854. — Bull. geol. inst. Uppsala, no. 5, vol. XXXVI, pp. 35—49, pl. 1. Uppsala.

KIELAN, Z., 1957: On the trilobite family Staurocephalidae. — Acta paleont. Polonica, vol. II, nos. 2—3, pp. 155—180, pls. I—VI. Warszawa.

KIÆR, J., 1897: Faunistische Uebersicht der Etage 5 des norwegischen Silursystems. — Skr. Norske Vidensk.-Akad., I kl., 1897, no. 3, pp. 1—76. Kristiania (Oslo).

LINNARSSON, J. G. O., 1869: Om Vestergötlands Cambriska och Siluriska aflagringar. — Kungl. Svenska Vetensk.-Akad. Handl., 8, 2, pp. 1—89, pls. I—II. Stockholm.

MÄNNIL, R., 1958: Trilobity semejctv Cheiruridae i Encrinuridae iz Estonii. (English Summary: Trilobites of the families Cheiruridae and Encrinuridae from Estonia). — Ensv Teaduste Akadeemia Geoloogia Instituudi Uurimused, III, pp. 165—212, pls. I—VIII. Tallinn.

NIESZKOWSKI, J., 1857: Versuch einer Monographie der in den silurischen Schichten der Ostseeprovinzen Vorkommenden Trilobiten. — Dorpat, Arch. Nat. Liv. — Ehst- und Kurlands, S. 1, Bd. 1, pp. 517—626, pls. 1—3.

SCHMIDT, F., 1881: Revision der ostbaltischen silurischen Trilobiten. — Memoires de L'Akad. Imp. des Sciences de St.-Petersbourg, Ser. VII, Vol. XXX, No. 1, Ab. I, pp. 1—237, pls. I—XVI. St.-Petersbourg.

— 1907: Revision der ostbaltischen silurischen Trilobiten. — Memoires de l'Akad. Imp. des Sciences de St.-Petersbourg, Ser. VIII, Vol. XX, No. 8, Ab. VI, pp. I—XV, 1—104, pls. I—III. St.-Petersbourg.

STØRMER, L., 1940: Early descriptions of Norwegian trilobites. The type specimens of C. Boeck, M. Sars and M. Esmark. — Norsk geol. tidsskr., 20, pp. 113—151, pls. 1—3. Oslo.

— 1953: The Middle Ordovician of the Oslo Region, Norway. 1. Introduction to stratigraphy. — Norsk geol. tidsskr., 31, pp. 37—141. Bergen.

TRIPP, R. P., 1954: Caradocian trilobites from mudstones at Craighead Quarry, near Girvan, Ayrshire. — Royal soc. of Edinburgh, Vol. LXII, Part. III, No. 16, pp. 655—693, pls. I—IV. Edinburgh.

TÖRNQUIST, S. L., 1884: Undersökningar öfver Siljansområdets trilobitfauna. — Sveriges geol. unders., Ser. C, No. 66, pp. 1—101, pls. I—III. Stockholm.

WARBURG, E., 1925: The trilobites of the Leptaena limestone in Dalarne. — Geol. Inst. Upsala, Bull., vol. 17, pp. 1—446, pls. 1—11. Uppsala.

WIGAND, G., 1888: Ueber die Trilobiten der silurischen Geschibe in Mecklenburg. — In Beitrag zur Geologie Mecklenburgs, Ab. 4, pp. 39—101, pls. VI—X. Rostock.

Manuscript received December 22, 1960.

Printed December 1961.

MIDDLE ORDOVICIAN CHEIRURID TRILOBITES FOUND IN NORWAY AND THEIR KNOWN STRATIGRAPHIC AND GEOGRAPHIC OCCURRENCE.

* = not previously recorded from Norway

	Horizon ? = horizon uncertain													Districts							
	4 aα	4 aβ	Ceph.	4 bα	Ech.	4 bβ	Coel.	4 bγ	Cycl.	4 bδ	Sphaer.	Gagnum Sh.	Mjøsa Lmst.	Skien–Langesund	Oslo–Asker	Ringerike	Hadeland	Toten	Nes–Hamar	Ringsaker	Biri
CHEIRURIDAE																					
*Paraceraurus helgoeyensis**	.	.	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	+	.	.
—, *perlongus*	.	.	.	.	.	.	.	?	.	.	.	.	.	.	+	.	.	.	.	.	.
*Cyrtometopella tumula**	.	+	.	+	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	.
— *askerensis**	.	+	.	.	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	.
*Nieszkowskia norvegica**	.	+	.	.	+	.	.	.	.	.	.	.	.	+	+	.	.	.	.	.	.
*Hemisphaerocoryphe inflatus**	.	.	.	+	.	.	+	.	.	.	.	.	.	.	+	.	.	+	.	.	.
— *sphaericus*	.	+	.	+	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	.
PLIOMERIDAE																					
Pliomerops sp.	+	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	+
ENCRINURIDAE																					
Encrinurus sp.*	.	.	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	+	.	.
Atractopyge brevicauda	.	.	.	.	.	.	.	.	.	+	.	+	+	.	.	+	+	.	+	.	.
— *dentata*	+	+	+	+	+	+	+	.	.	.	.	.	.	+	+	+	+	.	+	.	.
— *gracilis**	.	.	.	.	.	.	.	.	.	+	+	.	.	.	+	.	+	.	.	.	.
— *kutorgae**	.	.	.	+	.	.	.	.	+	.	.	.	.	.	+	.	.	.	.	+	.
— aff. *kutorgae**	.	.	.	.	.	.	+	.	.	.	.	.	.	.	.	.	+	.	.	.	.
— sp.	.	+	.	.	.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	.	.
Staurocephalus sp.	.	.	.	.	.	.	.	.	.	+	.	.	.	.	+	.	.	.	.	.	.

Ceph. = Cephalopod Shale (4aβ?) Ech. = Echinosphaerites Beds (4bα?). Coel. = Coelosphaeridium Beds (4bβ?). Cycl. = Cyclocrinus Beds (4bγ?). Sphaer. = Sphaeronid Limestone (4bδ?)

PLATE 1

The photographs are not retouched, but the specimens were whitened with ammonium chloride before photographing.

Paraceraurus helgoeyensis n. sp. — p. 283.

Fig. 1. x 2.0. Holotype cranidium. P.M.O. no. 67937. Mjøsa Limestone (4bδ). Bergevika, Helgøya. Coll.: G. Henningsmoen, 1954.
Fig. 2. x 1.9. Pygidium. P.M.O. no. 67981. Mjøsa Limestone (4bδ). Bergevika, Helgøya. Coll.: G. Henningsmoen, 1954.
Fig. 3. x 2.0. Hypostome. P.M.O. no. 68010. Mjøsa Limestone (4bδ). Bergevika, Helgøya. Coll.: G. Henningsmoen, 1954.

Paraceraurus perlongus (BRØGGER, 1882) — p. 286.

Fig. 4. x 2.0. Holotype (by monotypy) cranidium. P.M.O. no. H 2636. Probably 4bγ. Oslo. Coll.: ?

Cyrtometopella askerensis n. gen., n. sp. — p. 289.

Fig. 5. x 3.5. Holotype cephalon. P.M.O. no. 72226. 4aβ. Terneskjær, Asker. Coll.: N. Spjeldnæs, 1953.
Fig. 6. x 4.0. Right side view of cranidium, showing the convexity of the glabella. P.M.O. no. 72224. 4aβ. Terneskjær, Asker. Coll.: N. Spjeldnæs, 1953.

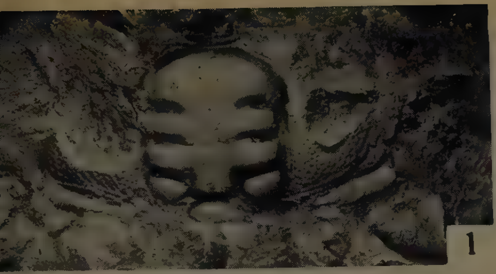
Cyrtometopella tumula n. gen., n. sp. — p. 288.

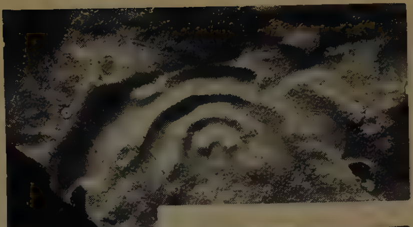
Fig. 7. x 4.35. Holotype cranidium. P.M.O. no. 72146. Transitional beds between 4aβ and 4bα. SW side of Ildjernet, Asker. Coll.: N. Spjeldnæs, 1957.
Fig. 8. x 4.0. Right side view of the specimen in fig. 7.
Fig. 9. x 2.0. Dorso-ventrally pressed pygidium. P.M.O. no. 18910. 4bα. Drammensveien at Slependen, Asker. Coll.: A. Heintz, 1933—34.
Fig. 10. x 2.0. Laterally compressed pygidium showing the long pleural spines. P.M.O. no. 72149. Transitional beds between 4aβ and 4bα. SW-side of Ildjernet, Asker. Coll.: N. Spjeldnæs, 1957.

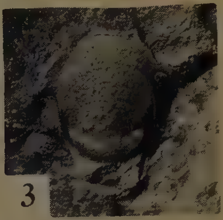
Hemisphaerocoryphe inflatus n. sp. — p. 292.

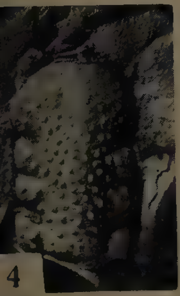
Fig. 11. x 2.0. Holotype cranidium. P.M.O. no. 67357. Coel. (4bβ?). Sund, Eina, Toten. Coll.: S. Skjeseth, 1951.
Fig. 12. x 2.0. Left side view of the specimen in fig. 10.

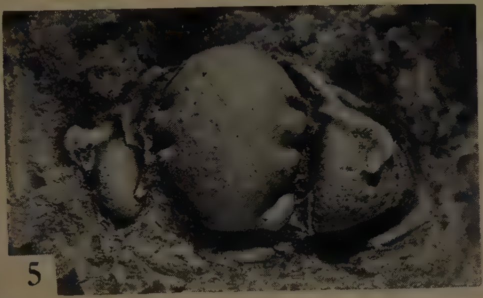
PLATE 1.


1

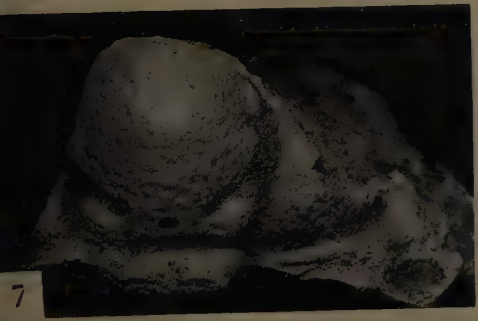

2

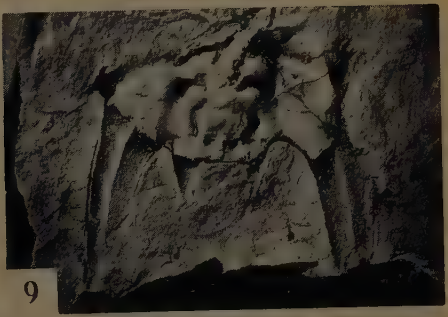

3

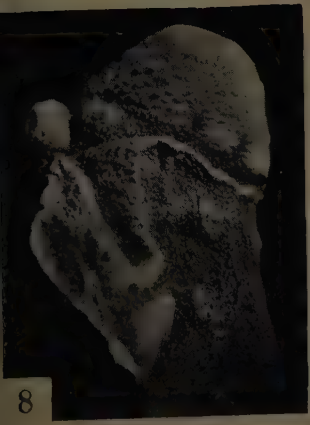

4

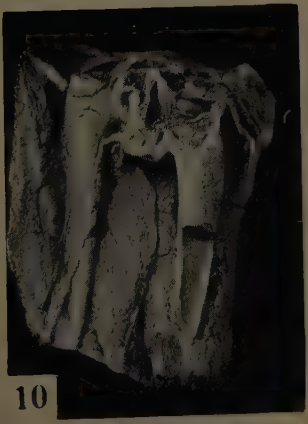

5

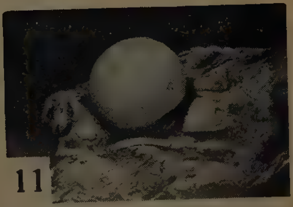
6


7


9


8


10


11

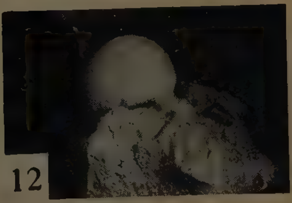

12

PLATE 2

The photographs are not retouched, but the specimens were whitened with ammonium chloride before photographing.

Nieszkowskia norvegica n. sp. — p. 291.

Fig. 1. x 2.0. Holotype cranidium. P.M.O. no. 8675. Ech. (4b*a*?). Skjellbukta, Frierfjord. Coll.: W. C. Brøgger, 1881.
Fig. 2. x 2.0. Right side view of the specimen in fig. 1.

Hemisphaerocoryphe sphaericus (ESMARK, 1883) — p. 293.

Fig. 3. x 2.0. Neotype cranidium. P.M.O. no. 61747. 4aβ or 4b*a*. Løken, Bygdøy, Oslo. Coll.: ?.

Pliomerops sp. — p. 295.

Fig. 4. x 2.0. Incomplete cranidium. P.M.O. no. 36600. 4a*a*. North of Redalen, Biri. Coll.: T. Strand, 1926.

Encrinurus sp. — p. 296.

Fig. 5. x 4.0. Pygidium. P.M.O. no. 72150. Mjøsa Limestone (4bδ). Bergevika, Helgøya. Coll.: N. Spjeldnæs, 1959.

Atractopyge brevicauda (ANGELIN, 1854) — p. 296.

Fig. 6. x 2.0. Cranidium. P.M.O. no. 67922. Mjøsa Limestone (4bδ). Bergevika, Helgøya. Coll.: G. Henningsmoen, 1954.
Fig. 7. x 1.9. Pygidium. P.M.O. no. 7384. 4bδ. Røysetangen, Ringerike. Coll.: J. Kiær, 1913.

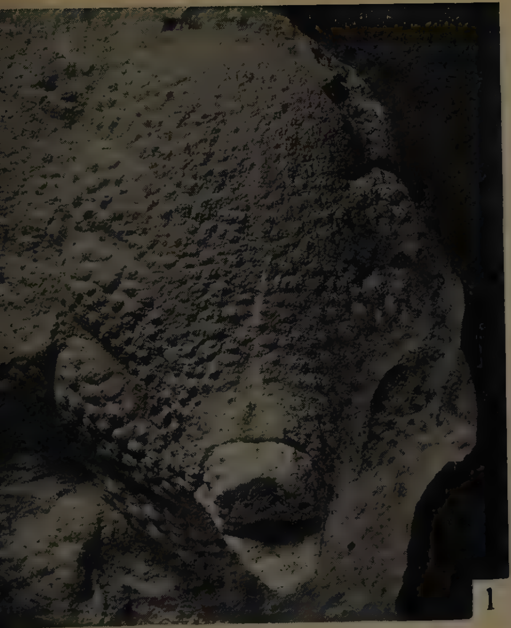
Atractopyge kutorgae (SCHMIDT, 1881) — p. 304.

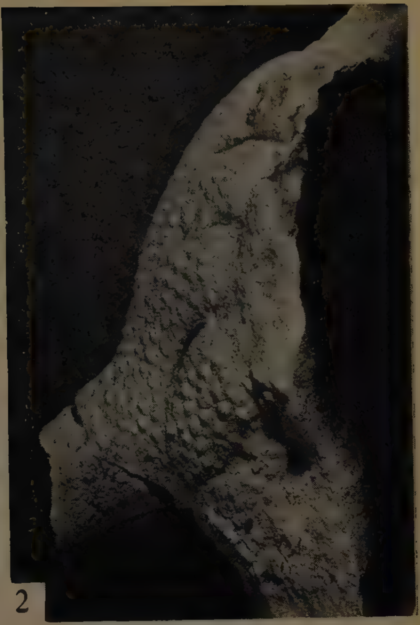
Fig. 8. x 2.0. Latex cast of cranidium showing test surface. P.M.O. no. 72152. 4b*a*. Bernhard Herres vei, Oslo. Coll.: N. Spjeldnæs, 1950.
Fig. 9. x 2.0. Exfoliated cranidium. P.M.O. no. 34241. Cycl. (4bγ?). South of Røykenvik, Ringsaker. Coll.: O. Holtedahl, ?.

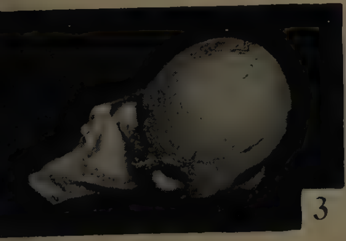
Atractopyge aff. *kutorgae* (SCHMIDT, 1881) — p. 305.

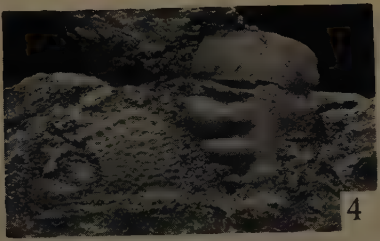
Fig. 10. x. 2.0. Pygidium. P.M.O. no. 2196. Coel. (4bβ?). Furuberget, Nes—Hamar. Coll.: O. Holtedahl, 1907.

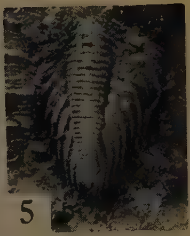
PLATE 2.

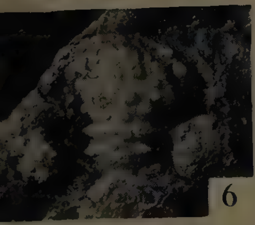

1

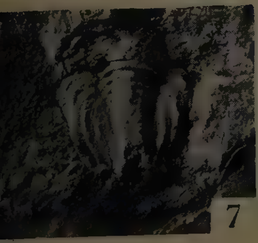

2

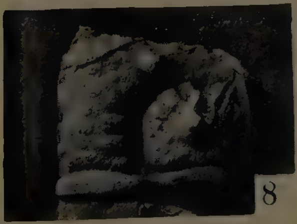

3

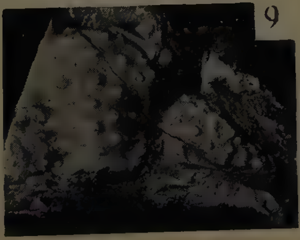

4


5


6


7


8


9

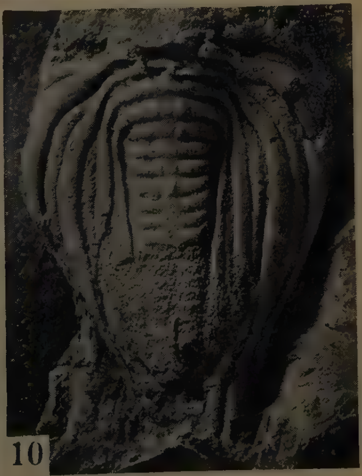

10

PLATE 3

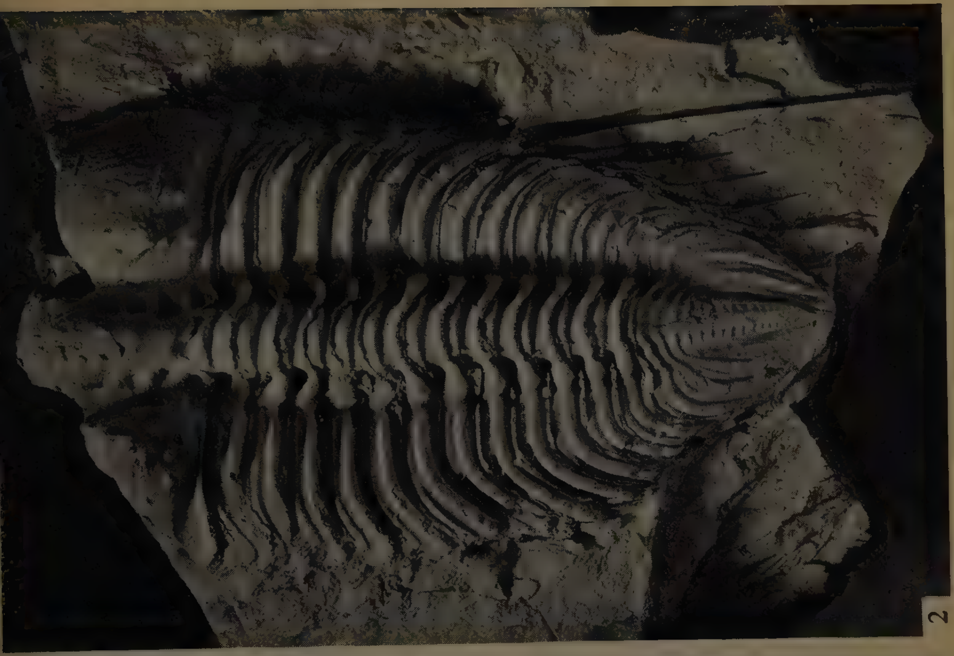
The photographs are not retouched, but the specimens were whitened with ammonium chloride before photographing.

Atractopyge dentata (ESMARK, 1833) — p. 298.

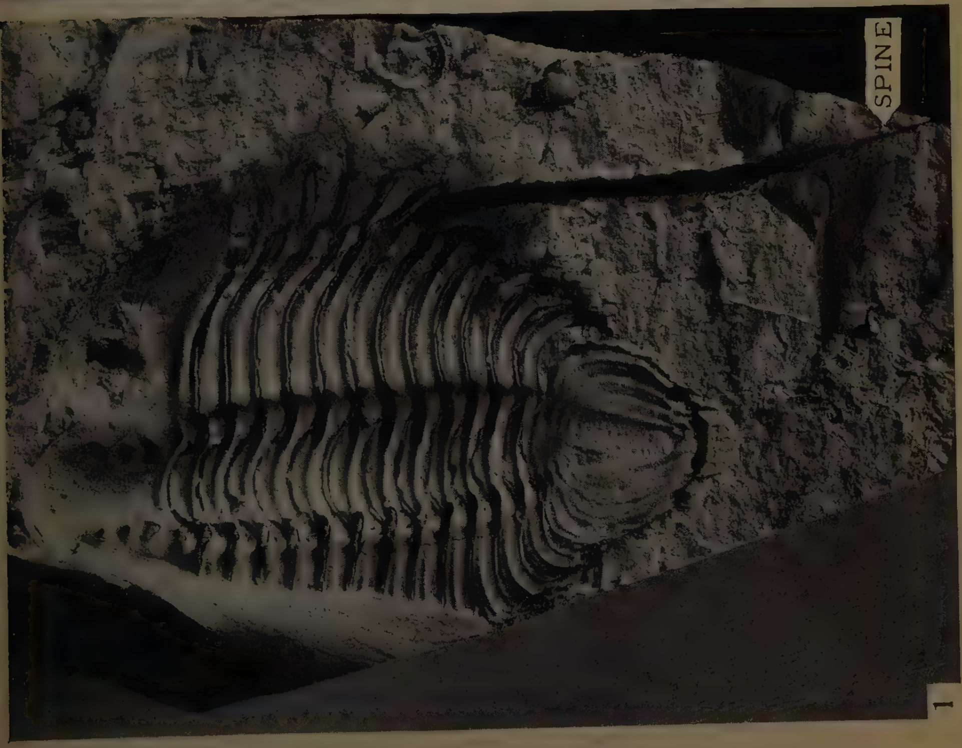
Fig. 1. x 1.7. Nearly complete dorso-ventral pressed specimen showing the spine on the 6. segment. P.M.O. no. 72151. 4b*a*. Husebybakken, Oslo. Coll.: J. Mauritz and H. Bjerkholt, 1958.

Fig. 2. x 1.7. Nearly complete specimen. P.M.O. no. 61609. 4b*a*. Railway section at Sinsen, Oslo. Coll.: O. Johansen, 1900.

PLATE 3.



2



1

PLATE 4

The photographs are not retouched, but the specimens were whitened with ammonium chloride before photographing.

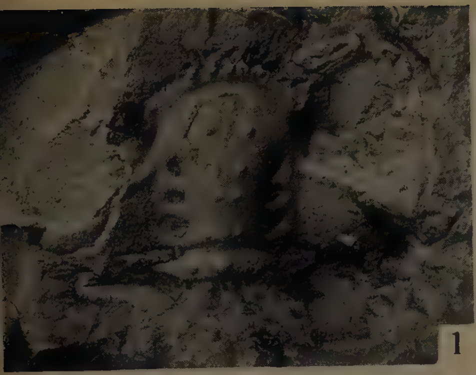
Atractopyge dentata (ESMARK, 1833) — p. 298.

Fig. 1. x 2.0. Cranidium showing the eye ridges and the spines in front of the glabella. P.M.O. no. 20366. 4b*α*. Delebukta, Asker. Coll.: J. Kiær, 1898.
Fig. 2. x 2.0. Librigena showing the stalked eye. P.M.O. no. 5277. Probably 4b*α*. Royal Castle, Oslo. Coll.: B. Esmark, ?.
Fig. 3. x 2.0. Pygidium. P.M.O. no. 6346. 4b*α*. Bratterud, Ringerike. Coll.: J. Kiær, 1913.
Fig. 4. x 2.0. Latex cast of lectotype pygidium. P.M.O. no. 56161. Probably 4b*α*. Royal Castle, Oslo. Old coll.

Atractopyge gracilis n. sp. — p. 302.

Fig. 5. x 2.0. Holotype cranidium. P.M.O. no. 6531. 4b*δ*. Terneholmen, Asker. Coll.: J. Kiær, exc., 1922.
Fig. 6. x 2.0. Cranidium. P.M.O. no. 5600. 4b*δ*. Terneholmen, Asker. Coll.: J. Kiær, 1921.
Fig. 7. x 2.0. Right side view of the specimen in fig. 6, showing the genal spines.
Fig. 8. x 2.0. Latex cast of librigena. P.M.O. no. 72145. 4b*δ*. Terneholmen, Asker. Coll.: N. Spjeldnæs, 1951.
Fig. 9. x 4.0. Pygidium. P.M.O. no. 9095. 4b*δ*. Terneholmen, Asker. Coll.: J. Kiær, 1921.

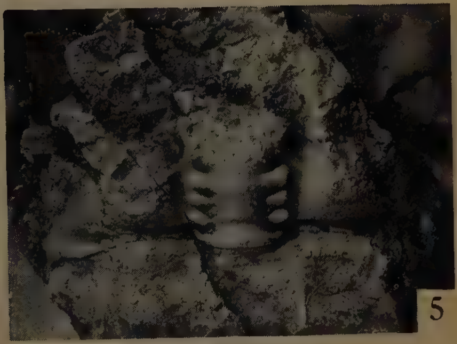
PLATE 4.


1

2

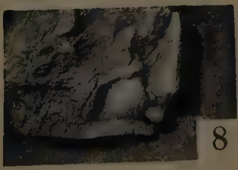
3

4


5

6

7


8

9

CONTRIBUTION TO THE MINERALOGY OF NORWAY

No. 11. On lanthanite in Norway

By

Per Chr. Sæbø
(Norges Geologiske undersøkelse).

Abstract. The first occurrence of lanthanite in Norway is reported. Physical and optical properties, and semi-quantitative and qualitative spectrographic determinations of the rare earth elements are given. Comments on the distribution of the rare earths in allanite and lanthanite are given.

Introduction

Lanthanite $(La_1Ce)_2\ (CO_3)_3 \cdot 8\ H_2O$ was first discovered at the famous mineral deposits at Bastnäs, Sweden, about 1820, by Berzelius (1825). The mineral occurred as an alteration product of cerite. The chemical composition was not well known until the mineral was found at Betlehem, Lehigh Co., Pa., U.S.A., in oxidized zinc ores. Good analyses that established the chemical composition were published by Smith (1854) and Genth (1857) on material from Betlehem, and Lindström (1910) on material from Bastnäs. Additional literature from before 1900 will not be reviewed here.

According to Palache, Berman and Frondel (1951), the mineral has also been reported as an alteration product after allanite in pegmatite at Baringer Hill, Llano Co., Texas, and in a magnetite deposit, Sandford Mine, Essex County, New York.

More recently lanthanite has been found in some unique parageneses in the Bearpaw Mountains, Montana, U.S.A. (Pecora & Kerr,

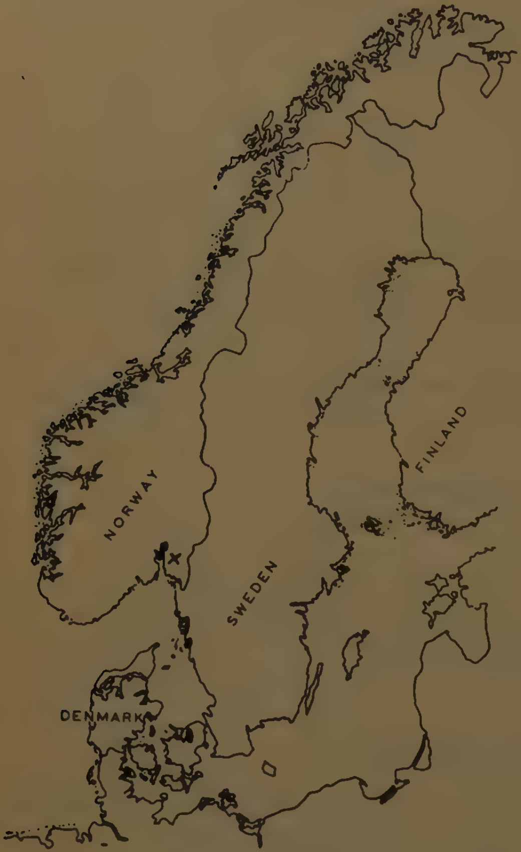


Fig. 1.

Map showing situation of the lanthanite locality. x.

1953). Here lanthanite occurs together with the new, closely related, rare earth carbonate calkinsite, $(La_1Ce)_2\ (CO_3)_3 \cdot 4\ H_2O$ both formed as late alteration products after the new, rare earth carbonate, burbankite and ancylite.

In 1957 Mr. P. Lie brought some samples to the Mineralogisk-Geologisk Museum from his granite pegmatite prospect, Sæteråsen, in the neighbourhood of his farm, Mjærum, in Hobøl, Østfold County, ca. 40 km south-east of Oslo. During examination of these samples under the binocular microscope I discovered extremely small amounts of lanthanite, growing as crystals in small druses in the quartz-feldspar mass.

Acknowledgements

I want to thank Prof. T. F. W. Barth for doing the first optical measurements on only three small crystal fragments. These observations led to the identification of the mineral.

I am indebted to Prof. I. Oftedal who kindly performed the spectrochemical determinations. Thanks are due to Dr. H. Neumann whose suggestions during our discussions were of great help. Thanks are also due to Prof. F. E. Wickmann, Stockholm, Sweden, who, on Dr. Neumann's request, promptly sent us a sample of the original lanthanite material from Bastnäs Sweden. Further I want to thank cand. mag. B. Nilssen for her contribution to this work.

Dr. P. H. Reitan kindly corrected the English. Mr. Brynhilsrud at the Geologisk Institutt, University of Oslo has taken the photographs of the lanthanite crystals. At last I want to express my gratitude to Mr. P. Lie for bringing this valuable material to the Mineralogisk-Geologisk Museum, University of Oslo, enabling me to make this contribution to the mineralogy of Norway.

The material

The samples were taken from the rather fine-grained border facies of the pegmatite. They consist of ordinary microcline, quartz, muscovite, and some 4—6 cm long crystals of allanite, the peripheral parts of which are always altered to a reddish, metamict substance, which is completely X-ray amorphous. The rock contains numerous very small druses, some of which are filled with pink, small, well-formed, tabular crystals of lanthanite, growing in irregular fan-shaped aggregates.

The optical measurements done by Prof. T. F. W. Barth indicated that the mineral was lanthanite. This was verified by comparison of the X-ray powder patterns of the mineral from Hobøl and the original lanthanite from Bastnäs, Sweden. (NEUMANN and BRYN, 1958).

Morphology of the lanthanite

The most detailed investigations of the forms and symmetry were done by BLAKE (1853) who established the rhombic symmetry of the mineral; VON LANG (1863) and FLINK (1910) calculated the axial

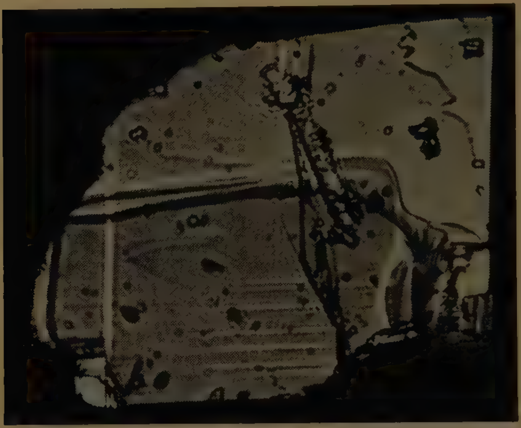


Fig. 2.
Fragment of a lanthanite crystal from Hobøl. Note the lamellar twinning along two directions parallel to the edge of (101). About 120x.

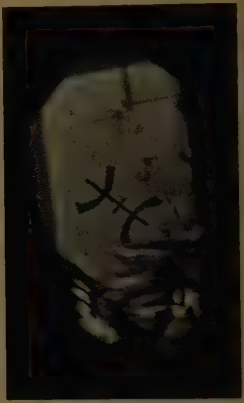


Fig. 3.
A typical well developed platy lanthanite crystal from Hobøl Cross-shaped twinning is present. The crystal lays on (010) and shows the forms {100}, {101}, {111} and {101} well developed. The optic orientation is given. The optic plan is parallel to (001). About 70 x.

ratios. Flink also described the twinning parallel to {101} and the characteristic lineations parallel the trace of {101} on {010}. Palache, Berman and Frondel (1951), based their description on the elements of the structural cell: a_o = 9.50 kX, b_o = 17.10 kX, c_o = 9.00 kX found by Wolfe and Frondel in 1947. (Stated in Palache, Berman and Frondel, 1951). This new orientation is used in this work[1]. Under the microscope the lanthanite from Hobøl shows the same forms as the mineral from Bastnäs and Bethlehem. The crystals are tabular parallell to {010}. The mineral also shows the faint lineation on {010}, parallel the edge of {101} reported by Flink (1910). This may represent normal submicroscopic twinning parallel {101}. Normal twinning parallel {101} is often found. The twins sometimes form cross-shaped intergrowths with each other as shown in figures 2 and 3. On account of the preferred orientation on (010), these twins resemble albite twins cut normal to the twin-plane.

[1] The old orientation was given by von Lang and Flink. Transformation: von Lang and Flink to Palache, Berman and Frondal (1951): 010/002/100.

Optical data

Determinations on many small crystals lead to the following values:

$\beta = 1.586 \pm 0.004$.

$\gamma = 1.615 \pm 0.005$. The optic plane is (001). The clevage parallel to (010) is well developed.

α was not measured on account of the pronounced preferred orientation of the crystals on (010) in the immersion slides.

α probably does not deviate from the earlier published value (LARSEN and BERMAN, 1934) $\alpha = 1.520 \pm 0.01$ because the values measured for β and γ are in good agreement with those reported by LARSEN and BERMAN (1934).

The normal axial dispersion $r < v$, weak, given by LARSEN and BERMAN (1934) was not detected in the lanthanite from Hobøl. 2V (−) = 63 ± 1°, measured on the universal stage. This is in agreement with earlier published data on lanthanite.

Table 1.

Locality	2V (−)	Author
Bethlehem	62°	Larsen & Berman, 1934
?	62°	Winchell, 1956
Montana	60°	estimated; Pecora & Kerr, 1953
Hobøl	63 ±1°	Sæbø, 1960

The genesis of the lanthanite

Lanthanite formed as a late secondary mineral in small cavities in the surroundings of the altered allanite. The rare earths were evidently supplied from the decomposed allanite and brought into solution as hydroxides by the action of supergene solutions rich in alkalies and CO_2, In principle the mode of formation is similar to that found experimentally by MELLOR (1924). He synthesized lanthanite and probably also calkinsite by the reaction of La and Ce hydroxides with $NaHCO_3$ — solution saturated with CO_2. The precipitate crystallized by standing.

The lanthanite in nature probably precipitated from similar solutions, but also containing other elements, Al, Si, Fe, Ca, etc. The formation of lathanite took place at very low temperature.

Distribution of rare earths in allanite and lanthanite

Material from the core and the totally altered red coloured rim zone of allanite crystals was isolated for optical spectrograms and the results compared with that of the lanthanite.

Table 2.

Distribution of the elements in the allanite and lanthanite

Material	Major elements	Minor elements <1%	Trace
Allanite, core	Mg, Ca, Al, Si, Fe, Mn	Th, Y, Ti	Yb, Pb, Be, V
Allanite, rim..........	Mg, Ca, Al, Si, Fe	Mn, Ti	Th, Y, Pb, V
Lanthanite	La		Y

Determinations by Prof. I. Oftedal.

Only very small amounts of material was used for the spectrograms. Ce has not been sought for.

The results, however, might indicate that La and also Ce are easely precipitated as lanthanite. Most of the other RE are probably carried away by the supergene, altering solutions.

Synthetical work on calkinsite and lanthanite

Cand. mag. B. Nilssen at Mineralogisk Museum, University of Oslo, carried out a few experiments in order to synthesize lanthanite. The procedure given by Mellor (1924) was used. The precipitates were examined by the X-ray powder method. These experiments showed that calkinsite forms much more readily than lanthanite. Only by very careful drying of the crystallized precipitates, B. Nilssen was able to produce a mixture of calkinsite and lanthanite.

These experiments show clearly that Mellor and co-workers probably obtained a calkinsite-lanthanite mixture rather than pure lanthanite. No further work on the phase relationships between calkinsite and lanthanite has been carried out.

REFERENCES

BERZELIUS (1825): Kohlensaures Cereroxydul. Zeitschr. f. Min. Vol. 2. p. 209.

BLAKE, W. P. (1853): On the Occurrence of crystallized Carbonate of Lanthanum. Am. J. Sc. Vol. 16 p. 228—230.

FLINK, G. (1910): Bidrag til Sveriges mineralogi. II. Ark. Kjemi, Vol. 3. p. 163—166.

GENTH, F. A. (1857): Contribution to Mineralogy. 14. Lanthanite. Am. J. Sc. Vol. 23. p. 245.

VON LANG, V. (1863): On the Crystalline Form of Lanthanite. Phil. Mag. and Journ. Sc. Vol. 25 Fourth Series p. 43.

LARSEN, E. S. and BERMAN, H. (19340): The Microscopic Determination of the nonopaque Minerals. Geol. Survey Bull. 848. Wash. D.C.

LINDSTRÖM, G. (1910): Om lanthanite. Geol. För. Förh. Vol. 32. p. 206—214.

MELLOR (1924): A Comprehensive Treatise on Inorganic and Theoretical Chemistry. (1922—1937). Vol. 5. p. 664.

NEUMANN H. and BRYN, K. Ø (1958): X-ray powder patterns for mineral identification. IV Carbonates. Vid. Akad. Avh. I Mat.-Naturvit. kl. 1958. No. 1, Oslo.

PALACHE, C., BERMAN, H. and FRONDEL, C. (1951): The system of mineralogy. Vol. II. 7. ed., John Wiley & Sons, Inc. New York, 132. p. 241—243.

PECORA, W. T. and KERR, JOE, H. (1953): Burbankite and calkinsite, two new carbonate minerals from Montana. Am. Min. Vol. 38. p. 1169—1183.

SMITH, J. LAW. (1854): Reexamination of America Minerals. 42. Lanthanite. Am. J. Sc. Vol. 18. p. 378.

WINCHELL, A. N. and WINCHELL, H. (1959): Elements of Optical Mineralogy, Part II. Fourth ed. John Wiley & Sons, Inc. New York, third printing. 1959.

Manuscript received December 2, 1960.

Printed December 1961.

NOTISER

Refugieproblemet og de kvartærgeologiske metodene

Av

IVAN TH. ROSENQVIST

I publikasjonen «Svensk Naturvetenskap», Fjortonde årgång, utgitt av Statens Naturvetenskapliga Forskningsråd, Stockholm, har Eilif Dahl en artikkel med ovenstående titel. Artikkelen består i det vesentlige av forskjellige motargumenter mot de forskere som hevder eller har hevdet at nedisingen under siste istids maksimum dekket Fennoskandia i alle fall slik at det ikke *kan påvises* områder hvor noe større antall planter eller dyr kunne leve. Dahls artikkel er for en stor del basert på kvartærgeologiske og mineralogiske argumenter. Særlig synes han å legge stor vekt på studiet av skiktgittermineralene. Det Svenska Naturvetenskapliga Forskningsråds årbok er ikke en publikasjon som egner seg for en videregående vitenskapelig diskusjon. Dahls artikkel bør imidlertid ikke stå uimotsagt, særlig på grunn av at hans kategoriske utsagn kan få lesere som ikke spesielt har arbeidet med disse problemer til å tro at Dahls oppfatning er alminnelig anerkjent.

Jeg vil ikke her ta opp de kvartærgeologiske argumenter, men skal holde meg til de kjemisk-mineralogiske.

Dahl insinuerer flere ganger at de norske forskere som tidligere har bearbeidet disse problemer, f. eks. Hans Holtedahl, Johan Moum og jeg selv, ikke har vært tilstrekkelig kritiske. Det er mulig at våre studier virker på en slik måte på en botaniker. Imidlertid kan jeg berolige Dahl og lesere av hans artikkel med at de mineraloger og geologer som har behandlet dette problemkompleks har vært klar over begrensningen og mulighetene ved de arbeidsmetoder som har vært anvendt, noe Dahl ikke synes å være. Det er f. eks. en misforståelse fra Dahls side når han tror at Moums og min (1955 og 1957) påvisning av montmorillonitt som sekundært forvitringsmineral i topplagene på de sydøstnorske leirer bare gjelder *en* leire nær Oslo, og at «det kan rettes alvorlige innvendinger mot identifikasjonen», som bygger på «en enkelt svak linje i et røntgendiagram av et glyserolpreparat, men det er ikke undersøkt om den samme linje opptrer i uforvitret leire». Det dreier seg her om et rent generelt fenomen. Det forhold som ble påpekt i det arbeid Dahl åpenbart henviser til, er at det i den geologisk sett meget korte tid som er gått etter at erosjon hadde skapt de nåværende overflateformer i våre marine leirer, er inntrådt en betydelig mineralogisk endring i topplagene, dvs. ned til noen få meter, og at disse deler av leirprofilene atskiller seg på en signifikant måte fra de under-liggende leirer. I virkeligheten er det forhold som Odd Gjems (1960) har funnet, nemlig montmorillonitt som sekundært mineral i fraksjoner under 20 mikron i A_2-skiktet i resente podsol-profiler, nettopp en del av det samme generelle bilde som Moum og jeg fremla 6 år tidligere, og som Dahl prøver å fralegge beviskraft.

Når det gjelder forvitringsdybde som funksjon av tid, så avhenger, som også Dahl sier, dette selvfølgelig av svært mange faktorer, hvoriblant bergartenes permeabilitet og diffusjonskoeffisienten for oppløste ioner i porevannet bør regnes blant de viktigste. Permeabilitetskoeffisienten i tilsynelatende meget like bergarter som f. eks. de grå Båhuslen-Iddefjord granitter eller de massive Trondhjemitter har vi funnet kan variere innen meget vide grenser, $K = 10^{-7} - 10^{-9}$cm.sek.$^{-1}$ uten at man kan se dette på håndstykke eller tynnslip før målinger er tatt. Diffusjonskoeffisienten for hydrogen og oppløste ioner kan på samme måte variere flere tierpotenser i tilsynelatende like bergarter. De kjemiske prosesser som fører til dannelse av hydroglimmer, vermikulitt og montmorillonittmineraler kan derfor ha vesentlig forskjellig hastighet i bergarter som man ved en overfladisk betraktning ville anta var helt like.

Dahl argumenterer med at tiden siden istiden er for kort til at kjemisk forvitring kan ha inntrådt. Her beror hans oppfatning på en misforståelse. I et møte som ble holdt i Cambridge april 1961 for å diskutere de geokjemiske effekter av forvitring, viste R. Glentworth (1961) at de underjordiske deler av granittblokker i en stensirkel i Aberdeenshire hadde undergått kjemisk forvitring til en dybde av ca. 3” siden Bronsealderen. Den kjemiske forvitring er under våre klimatiske betingelser og under gunstige forhold vesentlig raskere enn Dahl synes å tro. På den annen side må man være oppmerksom på at såvidt forvitringsprosessene er diffusjonsbetinget vil forvitringsdybden være en funksjon av kvadratroten av tiden. Den blir derfor ikke mer enn 3—4 ganger mer gjennomtrengende om man forutsetter at alderen er fra nestsiste istid enn om man forutsetter at det er startet etter siste istids maksimum. Aktiveringsenergien for de forskjellige geologiske diffusjonsprosesser varierer ganske betydelig, men vil ved forvitring av bergarter i det vesentlige være avhengig av aktiveringsenergiene for diffusjon i væskefasen. Disse ligger stort sett mellom 4000 og 11000 cal. pr. mol. Dette vil igjen si at temperaturavhengigheten ikke er så stor at det kan føre til størrelsesordeners forskjell i vitringshastigheten innen de temperaturintervaller vi her har å regne med.

Til sist har vi punktet om hvorvidt forvitret berg kan tenkes å bli bevart på tross av at området har gjennomlevet en nedising. Her i landet er dette påvist på mange steder, f.eks. av Tom Barth (1939), og under det ovennevnte møte i Cambridge viste D. F. Ball (1961) at man på øya Rhum hadde en 50 fot dypt forvitret gabbro under morene. Denne forvitringssone representerer i toppen et interglacialt podsol-profil fra B-horisonten og nedover. Isen greide her altså bare å fjerne A_1 og A_2 horisontene.

For studiet av refugieproblemet må mange metoder tas i bruk. Biologene har sine metoder. Innen de krystallkjemiske/mineralogiske fagområder må mange arbeidsmetoder sammenholdes, og man må forstå metodenes betydning.

De biologiske og særlig botaniske argumenter som er ført i marken i forbindelse med overvintringsteorien hører til Dahls fagområde, der jeg savner forutsetninger for diskusjon.

HENVISNINGER

DAHL, EILIF (1961) Refugieproblemet og de kvartærgeologiske metodene. Svensk Naturvetenskap fjortonde årgång, Statens Naturvetenskapliga Forskningsråds årsbok, Stockholm 1961.

MOUM, J. og ROSENQVIST, I. TH. (1955) Kjemisk bergartsforvitring belyst ved en del leirprofiler. Norsk Geol. Tidsskr., Nr. 34, Bergen.

— (1957) On the weathering of young marine clay. Norwegian contrib. to the 4th Int. Conf. on Soil Mechanics and Foundation Engineering, No. Ia/18, pp. 5—7, London.

GJEMS, ODD, (1960) Clay minerals in podsol profiles in Fennoscandia. Clay Min. Bull. Vol. 4, No. 24. Galashiels.

GLENTWORTH, R. (1961) (Se referat Nature June 17, 1961, p. 1067).

BARTH, T. F. W. (1939) Norske mineraler av beidelittgruppen. Norsk Geologisk Tidsskrift Vol. 19, pp. 300—310.

BALL, D. F. (1961) (Se referat Nature June 17, 1961, p. 1067).

ON SUB SOLIDUS RELATIONS OF SILICATES

II. Preliminary Note on the Chemistry of Melilites

By

Olav H. J. Christie.
(Mineralogisk-Geologisk Museum, Oslo)

Raaz (1930) and Warren (1930) independently first described the atomic structure of melilite. Smith (1953) re-examined the structure using a natural crystal from the same locality as that of Warren's material.

The melilite structure is a sheet structure and therefore melilites are phyllosilicates and not sorosilicates as pointed out e.g. by Strunz (1957). The sheets of the melilite structure consist of MgO_4, $A1O_4$ and SiO_4 tetraheda having corners in common. The sheets are held together by Ca-O bonds. Some sets of the Si-O and of the Ca-O bonds are so long that they must be considered to have a marked homopolar character. (Smith (1953)).

As will be shown in a later paper Ca may be substituted by Na in melilites if this substitution is followed by a substitution of F for O. The location of the F atoms in the structure is not known, but there may be an analogy between the homopolar Ca-O bond in melilite and the Ca-OH bond in afwillite (Megaw (1952)). In afwillite Si occurs inside isolated tetrahedra. The tetrahedra are connected by Ca-O bonds, and the polyhedra around the Ca atoms are distorted (as they are in melilites). Megaw has shown that some of the corners in the Si tetrahedra are occupied by OH groups instead of by O and the corresponding Ca-OH distances are greater than are the Ca-O distances.

A similar explanation may be used to locate the F ions in artificial NaF-melilites. The unshared corners of the SiO_4 tetrahedra in the melilite sheet have, according to Smith (1953) insufficient Pauling valence bonds to justify the location of an O'' ion. It is probable, therefore, that the unshared corners of the SiO_4 tetrahedra are occupied by F in NaF-melilites and by OH groups in NaOH-melilites.

Among the attempts to produce melilites of the general formula $A_2BC_2O_7$ many were unsuccessful by apparently unexplainable reasons. Therefore, this formula should be revised. The oxygens of the unshared corners of the SiO_4 tetrahedra are structurally different from the oxygens shared between two tetrahedra since their distance from the metal atoms indicate that the bond is not purely ionic. The present author suggests that the general formula of melilites should be written A_2X_2 (BC_2Y_5) where the bonds of homopolar character are formed by the A and X atoms and the rest of the sheet is made up by the BC_2Y_5 part of the formula.

Empirically this formula has the great advantage that it permits the construction of strict rules for the formation of melilites. It seems to be a condition for the formation of melilites that the charge of the BC_2Y_5 part of the formula is small and definitely lower than one per mole. The present author will return to this in a more detailed paper which is now in preparation.

The two most important melilites are åkermanite, Ca_2O_2 $MgSi_2O_5$, and gehlenite, Ca_2O_2 $AlSiAlO_5$. Natural melilites normally contain some sodium and potassium and many attempts have been made to explain the mechanism of this substitution.

Schaller (1916) introduced the Na-sarcolite molecule, $Na_6Al_2Si_3O_{12}$, as a melilite end member and Buddington (1922) showed that this molecule, if mixed in the proportion 1:9 with the $Ca_3Al_2Si_3O_{12}$ molecule, is taken into complete solid solution in melilites. WINCHELL (1924), however, doubted this because the stochiometry of the Na-sarcolite molecule is different from that of melilite, and he suggested that a $Na_2Si_3O_7$ melilite existed. However, GOLDSMITH (1948) showed that this molecule is not taken into solid solution at all in åkermanite, and that it is accepted in the gehlenite molecule only to an extent of 15 percent by weight. These values are too low to account for the alkali content of natural melilites, and Goldsmith suggested that the substitution of Na for Ca in melilites had to be followed by a substitution of OH for O. Later SMITH (1953) showed that there is no relation between the water content and the alkali content of natural melilites. Thus, many conflicting ideas have been offered to explain the substitution mechanism leading to the formation of natural alkali melilites.

The present author attacked this problem by means of mineral synthesis, and demonstrated that two different mechanisms may lead to the formation of alkali melilites:

1) Gehlenite-containing melilites take some $Na_2Si_3O_7$ and some $K_2Si_3O_7$ in solid solution in amounts depending upon the gehlenite content as well as upon the temperature. When $Na_2Si_3O_7$ and $K_2Si_3O_7$ are taken in solid solution in gehlenite, the BC_2Y_5 part of the general formula becomes charged. At 1300° C the maximum charge of the BC_2Y_5 part seems to be $+ 0.3$ per miole. At 1000° C it is around $+0.2$ per mole.

2) The Ca-O bonds of gehlenite and åkermanite may be changed by Na-F, to a certain extent also by K-F bonds. By this substitution mechanism the BC_2Y_5 part of the general formula remains unchanged. No attempts have been made by the present author to produce the OH analogues of the fluorine melilites, but there are no structural restrictions to the formation of such analoges.

One important melilite end member (Ca_2O_2, Na_2F_2) $SiBeSiO_5$, has been made artificially by the present author. Leukophane and melinophane are mixed crystals between this melilite and åkermanite and gehlenite respectively.

REFERENCES

Buddington, A. F. (1922): On some natural and synthetic melilites. Am. Jour. Sci. vol. 3 (1922) p. 35—87.

Goldsmith, J. R. (1948): Some melilite solid solutions. Jour. Geol. vol. 56 (1948) p. 437—447.

Megaw, Helen D. (1952): The structure of afwillite, Ca (SiO OH) .2H . Acta Kryst. vol. 5 (1952) p. 477—491.

Raaz, F. (1930): Die Struktur des Meliliths. Akad. Wiss. Wien 139 (1930) p. 465.

Schaller, W. T. (1916): Mineralogic notes. Nova Scotia Geol. Survey Bull. 610. 3d ser. p. 109—128.

Smith, J. V. (1953): Reexamination of the crystal structure of melilite. Am. Min. 38 (1953) p. 643—661.

Strunz, H. (1957): Mineralogische Tabellen. Ed. Akad. Verlagsges. Geest & Portig K.-G. Leipzig 1957.

Warren, B. E. (1930): The structure of melilite $(Ca, Na)_2(Mg,Al)\ (Si,Al)_2O_7$ Zeitschr. Krist. vol. 74 (1930) p. 131—138.

Winchell, A. N. (1924): The composition of melilite. Am. Jour. Sci. 5th ser. vol. 8 (1924) p. 375—384.

Manuscript received May 29. 1961.
Printed December 1961.

International Association of Sedimentologi.

The general assembly of the Association at Copenhagen in 1960 decided to edit its own Bulletin of Sedimentology. The first number will appear in January 1962. The subscription price will be \$ 6.50 for members and \$ 9.— for non-members of the Association.

Persons who wish to become a member of the Association should inform the General Secretary, Prof. Dr. D. J. Doeglas, Postbus 37, Wageningen, Holland and send at the same time \$ 1.10 (or ƒ 4.— (Dutch Guilders) to the International Association of Sedimentology c/o Rotterdamsche Bank, Utrecht, Holland.

The next International Sedimentological Congress will be held in Belgium and Holland in 1963.

Ivar Hessland.

NY LITTERATUR

Jac. Berzelius Brev, utgivna av Kungl. Svenska Vetenskapsakademien. Supplement 3 ved Arne Holmberg, 1961. 78 s.

Professor Holtedahl har gjort meg oppmerksom på denne publikasjon, som har betydelig interesse for norske lesere. Del II, «Berzelius' brevväxling med Esmark far och son» (s. 36ff), inneholder 6 brev fra Jens Esmark og 1 fra Berzelius fra årene 1825—35, samt 11 brev fra Hans Morten Thrane Esmark (prest i Eidanger og ivrig mineralog) og 9 fra Berzelius fra årene 1827—47. Foruten å formidle en mer personlig kontakt med disse gamle herrer gir brevene bl. a. glimt fra den tidlige utforskning av Norges mineraler, særlig dem fra Larvik—Langesund-trakten.

I brev av 15. oktober 1825 fra Jens Esmark finner vi følgende: «Af nye Opdagelser her i Norge veed ieg intet uden, at der er opdaget en smuk rosenrød Lepidolit norden for Trondhiem, saasnart ieg kand faae noget deraf, vil jeg tage mig den Frihed, at sænde Dem Prøver deraf.» Det er en oppgave for nåtidens mineraloger å gjenoppdage denne forekomst, men det må innrømmes at holdepunktene er ytterst sparsomme.

Ivar Oftedal.

D. D. Bloss: An introduction to the Methods of Optical Crystallography. VII + 294 sider, 176 figurer (Holt, Rinehart and Winston, New York, 1961).

For alle som skal arbeide med polarisasjons-mikroskopet og vil sette seg inn i de metoder som brukes i optisk krystallografi, vil denne boken være til glede og nytte. Den byr på mange måter noe nytt og godt. Jeg kan anbefale den fremfor eldre lærebøker.

Teksten er skrevet for begynnere og gir en teoretisk og praktisk fremstilling av hvordan de optiske konstanter av krystaller kan bestemmes ved hjelp av polarisasjons-mirkoskopet og immersjons-oljer. Teorien er klart og nøyaktig bygd opp over *ett* grunnbegrep: den optiske indikatriks (uten å gå inn på bølgeflater eller andre optiske hjelpefigurer), en meget anbefalelsesverdig metode.

Det er også meget nyttig at bestemmelse av lysbrytningsindekser ved immersjonsoljer er nøye forklart, like til hvordan man kan gjøre bruk av de farvede dispersjonsrender på korngrensene. Et appendix inneholder bestemmelsestabeller, hvorved over 1 000 mineraler kan identifiseres ved sin lysbrytning. Derved behøver studenten ikke å bruke tre forskjellige bøker — én for teori, én for praktiske metoder og en tredje for bestemmelsestabeller — for å utføre sitt arbeid.

Boken er delt i elleve kapitler som ofte slutter enten med en sammenfatning eller med et meget nyttig avsnitt: Questions and Problems. Der er tre appendixer: Properties of ellipses; Determinative optical tables and charts; Recording data. Det er et godt register og en liste med litteraturreferanser.

Tom. F. W. Barth

Walter Fischer: Gesteins- und Lagerstättenbildung im Wandel der wissenschaftlichen Anschauung. VIII + 592 sider, 12 tavler, 12 figurer, 16 × 24 cm format. Personenregister, Ort- und Sachregister. Pris DM 84.00. (Schweizerbart, Stuttgart, 1961).

Dette er et usedvanlig verk og anbefales på det sterkeste for alle som er alvorlig interessert i bergarter, mineraler og malmer. Av spesiell nytte vil verket være for dem som underviser og veileder ved universiteter og høyskoler. La tittelen ikke skremme; den yter på ingen måte rettferdighet til det interessante og verdifulle innhold. Boken er nemlig bl.a. en meget god og fyldig moderne lærebok i petrografi, petrologi og malmgeologi; dette i tillegg til den underholdende og ofte spennende historiske beretning om tidligere generasjoners skiftende meninger og teorier. Klart og konsist gir boken således en oversikt over både vitenskapens historie og vitenskapens nyeste resultater og de teorier den hviler på.

De første 300 sider behandler «Die *Gesteinslehre*» — eruptiver, sedimenter og metamorfe bergarter fra de eldste tider til fyldestgjørende forklaringer av Bowens reaksjonsserier, Goldschmidts geokjemi, Eskolas facieslære og Wegmanns tektonikk. Den neste del behandler «*Die Lagerstättenkunde*». Et meget morsomt kapitel er «die Anfänge der modernen Lagerstättenkunde», hvor den interessante «Bergbüchlein» fra omkring 1500 utførlig beskrives. Her kunne jeg si at det vel burde være nevnt at *das Bergbüchlein* så vel som *das Probierbüchlein* er utgitt på engelsk av The American Institute of Mining and Metallurgical Engineers (1949), mesterlig oversatt av Sisco and Smith, og forsynt med technical annotations and historical notes, som gir et meget interessant innblikk i datidens kunnskaper.

Kapitlet om «die Entwicklung der Lagerstätten-systematik» er klart, konsist og noe av det beste som er skrevet på dette felt. Så følger en nøyaktig gjennomgåelse av «Lagerstätten der magmatischen Abfolge», «der sedimentären Abfolge» og til slutt «der metamorphen Abfolge». De forskjellige oppfatninger blir klart gjengitt, og der er fyldige diskusjoner om de geologiske relasjoner, om de fysikalsk-kjemiske faktorer, etc.; hele den moderne teori anvendt på dette meget kompliserte felt er på en beundringsverdig måte fremstillet og klargjort.

Tom. F. W. Barth

NORSK GEOLOGISK FORENING

Møte nr. 393, torsdag 2. mars 1961.

i Drammensveien 78, Oslo.

Til stede var 34 medlemmer og 6 gjester.
Det var sendt innbydelse til møtet til Norges geografiske oppmåling.
Referat av generalforsamling og møte 392 ble lest opp og godkjent.

Følgende nye medlemmer ble innvalgt:

479. Cand. real. Magne Gustavson, Norges geologiske undersøkelse, Oslo — N.V. Foreslått av H. Skålvoll og S. Bergstøl.
480. Mr. P. H. Banham, B. Sc., F.G.S., Dept. of Geology, University of Nottingham, England. Foreslått av F. M. Vokes og T. G. Elder.
481. Siv.ing. Arne H. Christensen, Flyvåpnets forsyningskommando, Kjeller. Foreslått av L. Lund og B. Bruun.
482. Cand. mag. Per Jørgensen, Geologisk Inst., Blindern.
483. Stud. real. Jens Olaf Englund, Geologisk Inst, Blindern, begge foreslått av G. Liestøl og T. Strand.

Det ble bekjentgjort at styret har godkjent opprettelsen av en Stratigrafisk gruppe av Norsk Geologisk Forening.

Aftenens foredragsholder var *Scott B. Smithson* som snakket om «Geological interpretation of gravity anomalies in Norway».

I diskusjonen etter foredraget deltok Chr. Oftedahl, Whitaker, Strand, G. Liestøl, Jelstrup, Falkum, Vokes, Rosendahl, Trovåg, Padget, Dons, Spjældnes og foredragsholderen.

Møte nr. 394, torsdag 13. april 1961.

i Drammensveien 78, Oslo.

Tilstede var 23 medlemmer og 6 gjester.
Referat fra møte 393 ble lest opp og godkjent.

Aftenens foredragsholder var *J. H. McD. Whitaker* som snakket om: «Palæogeography from primary sedimentary structures».

I diskusjonen etter foredraget deltok Størmer og foredragsholderen.

Referat fra ekskursjon til Gruaområdet den 28 5—61. Ledere var statsgeolog S. Skjeseth og cand.mag. H. Bollingberg.

Det var ca. 20 deltakere i ekskursjonen.

GEOLOGNYTT

Norges geologiske undersøkelse.

Cand. real. MAGNE GUSTAVSON tiltrådte som vitensk. assistent I 1. jan. 1961.

Statsgeolog II KARI EGEDE LARSSEN har permisjon fra 1. aug. 1961. Vikar cand. real. CHR. DICK THORKILDSEN, tiltrådt 1. sept. 1961.

Midl. vit. ass. PER CHR. SÆBØ har permisjon fra 1. aug. 1961. Vikar cand. real. JENS HYSINGJORD, tiltrådt 15. aug. 1961.

Universitetet i Oslo.

Institutt for geologi.

Professor OLAF HOLTEDAHL er i forbindelse med Nansenjubileet den 10. oktober 1961 blitt tildelt Framkomiteens Nansenbelønning.

Amanuensis ved instituttets kjemiske laboratorium cand. real. OLAF HAGBERG sluttet høsten 1960. Som ny amanuensis er ansatt mag. scient. MICAEL HIORTH MLADECK.

I et nyopprettet universitetslektorat knyttet til instituttet er ansatt cand. real. INGE BRYHNI.

Dr. philos NILS SPJELDNÆS er ansatt som førsteamanuensis ved avdeling C.

Vitenskapelig assistent ved avdeling B cand. real. JENS HYSINGJORD sluttet våren 1961. Ny assistent er cand. mag. PER CHR. SÆBØ.

Dr. C. A. NELSON, California, oppholder seg ved instituttet (avd. A) med et National Science Foundation Senior Postdoctoral Fellowship, for å arbeide med eokambriske og nærstående avleiringer.

Mineralogisk-geologisk museum.

Cand. real. JOHAN NATERSTAD sluttet som vitenskapelig assistent 1. juni 1961.

Cand. real. ARNE GRØNHAUG ble ansatt som vitenskapelig assistent fra 1. august 1961 og slutter pr. 1. januar 1962 for å tiltre stilling som geolog I ved Veglaboratoriet.

Følgende gjestende forskere arbeider ved museet:
SCOTT B. SMITHSON, M.A., USA.
THOMAS G. ELDER, B.Sc., Great Britain.
MICHAEL PETTERSSON, B.Sc. ,Great Britain.
NORMAN PAGE, B.A., USA.
ARTHUR SYLVESTER, B.A., USA.

Paleontologisk museum.

Dr. phil. NILS SPJELDNÆS fratrådte stillingen som førstekonservator 28. februar 1961 for å tiltre stillingen som førsteamanuensis ved Institutt for geologi, Universitetet i Oslo, 1. mars 1961.

Stud. real. FRANK NIKOLAISEN har fungert som midlertidig vitenskapelig assistent fra 1. mars 1961.

Førstekonservator GUNNAR HENNINGSMOEN ble valgt til formann i Stratigrafisk Gruppe av Norsk Geologisk Forening på konstituerende møte 2. mars 1961.

Gjestende forskere:

Dr. STANISŁAW ORŁOWSKI, Warszawa. Kambriske faunaer og kambrisk stratigrafi. (Juni-Oktober 1961.)

Dr. ELLIS L. YOCHELSON, U.S. Geological Survey, Washington, D.C. Ordoviciske gastropoder. (August 1961—Juni 1962.)

Universitetet i Bergen.

Geologisk institutt.

Cand. real. FINN J. SKJERLIE tiltrådte som amanuensis I 1. juni 1961.

Cand. real. INGE BRYHNI fratrådte som vit. ass. 31. juli 1961.

Cand. real. TORE TORSKE tiltrådte som vit. ass. 1. aug. 1961.

Cand. real. JOHAN NATERSTAD tiltrådte 1. juli 1961 som vit. ass. lønnet av N.A.V.F.

Jordskjelvstasjonen.

Forskningsstipendiat MARKVARD A. SELLEVOLL ble beskikket som dosent fra 1. jan. 1961.

Cand. real. ANDERS SØRNES tiltrådte som vit. ass. 1. jan. 1961.

Statens vegvesen.

Cand. real. CHR. C. GLEDITSCH er fratrådt stillingen som geolog 1 ved Veglaboratoriet. Vikarierende i geologstillingen er cand. mag. PER DUGSTAD.

Geofysisk malmleting.

Geolog I FINN J. SKJERLIE sluttet ved GM 31/5 1961.

I den ledige stilling etter Skjerlie er ansatt geolog SVERRE SVINNDAL fra 1/7 1961.

Eksamener.

Universitetet i Oslo.

Følgende har fullført embetseksamen med mineralogi-petrografi som hovedfag. Rekkefølgen er kronologisk.

SIGBJØRN JARLE KOLLUNG: «Metamorfe og eruptive bergarter på Hitra.»

BORGHILD NILSSEN: «Noen geologiske undersøkelser av Herefossgraniten.»

Per Ragnar Lund: «En geologisk undersøkelse på den sør-østre del av kartbladet Salangen.»

Boye Flood: «En undersøkelse av Straumfjellets kobberfelter og omliggende bergarter.»

Harald Øyen Skålvoll: «Kvartærgeologiske undersøkelser i herredene Romedal og Stange, Hedmark fylke.»

Haldis Johanne Bollingberg: «Spektrografisk undersøkelse av sporelementer i blyglans og sinkblende, med enkelte petrografiske og tektoniske iakttagelser fra Grua, Hadeland.

Følgende har fullført embetseksamen med historisk geologi og paleontologi som hovedfag.

Dag Arne Kjell Risdal: «En undersøkelse av kvartære, biostratigrafiske soner i Drammen på grunnlag av foraminiferer.»

Georg Hamar: «Stratigrafi og tektonikk i traktene omkring Norderhov, Ringerike.»

Universitetet i Bergen.

Embetseksamen med mineralogi-petrografi som hovedfag er fullført av:

Kjell A. Tjølsen: «En detaljert petrografisk-tektonisk undersøkelse av Ombo i Ryfylke.»

Johan Naterstad: «Noen geologiske undersøkelser i et profil langs Mundheimsdalen, Hordaland.»

Tore Torske: «Geologiske undersøkelser i området Mostadmarka—Selbustrand, Nord- og Sør-Trøndelag fylker.»

Embetseksamen med fysisk geografi som hovedfag er fullført av:

Frank Petter Myhre: «Om seismisk dispersjon.»

CIPEA.

Ved kongressen i København ble det valgt følgende Bureau for CIPEA — Comité Internationale Pour Etude des Argiles:

President: Professor I. Th. Rosenqvist, Oslo.
Sekretær: Dr. Poul Graff-Petersen, Købehavn.
Øvrige medlemmer: Professor C. W. Correnz, Tyskland.
Professor W. C. Dekeyser, Belgia.
Dr. R. C. Mackenzie, Storbritannia.
Professor J. Orcel, Frankrike.
Dr. Ralph Grim, USA.
Professor F. V. Chukhrov, USSR.

MEDLEMSLISTE

à jour november 1961,
[1]: korresponderende medlem.
*: livsvarig medlem.
Tallet i parentes er innvalgsåret.
(S), stifter (18. februar 1905).

*Adamson, Olge J., dr. Königlich Norwegische Botschaft, Drachenfelsstrasse 9, Bonn, Tyskland. (1945).
Ahlmann, Hans W:son, professor, Fregattvägen 6, Gräsdal, Stockholm S.V., Sverige. (1919).
Andersen, Bjørn, univ.stip. Geologisk institutt, Blindern, Oslo. (1952).
Asklund, Bror, professor, Sveriges Geologiska Undersökning, Stockholm 50, Sverige. (1938).
*Bache, Laura, lektor. Eiksbo, vei 509, Eiksmarka, Røa, Oslo. (1931).
Baker, Donald, B.Sc. The Ohio Oil Company, P. O. Box 269, Littleton, Co., U.S.A. (1951).
Banham, P. H., B. Sc., F.G.S. Department of Geology, University of Nottingham, England. (1961).
*Barth, Tom, F. W., professor. Geologisk museum, Oslo NØ. (1921).
Bergersen, Birger, professor. Anton Schjøtsgt. 13, Oslo (1921).
Bergstøl, Sveinung, cand. real. Geologisk museum, Oslo NØ. (1959).
*Berry, William B. N., professor. Dept. of Paleontology, University of California, Berkeley 4, Cal., U.S.A. (1960).
Bertheau-Hansen, Chr., bergingeniør. Holmenveien 52a, Smestad — Oslo (1942)
Birkeland, Tor, lektor. Avaldsnesgt. 36, Stavanger. (1946).
Birkenmajer, Krzysztof, dr. Laboratory of Geology, Polish Academy of Science, Slawkowska 17, Krakow, Polen. (1960).
Bjerrum, Lauritz, direktør. Norges geotekniske institutt, Blindern, Oslo. (1962).
*Bjørlykke, Harald, direktør. Norges geologiske undersøkelse, Trondheim. (1923)
Bollingberg, Halldis, cand. real., Mineralogisk Museum, Østervoldgade 7, København K., Danmark. (1960).
Boughton, Eric. Hay Tor, 921 Harrow Rd., Wembley, Middlesex, England. (1923).

*Broch, Olaf Anton, statsgeolog. Carl Kjelsens vei 25b, Sogn, Oslo. (1920).
*Bruun, B., ingeniør. Geologisk museum, Sarsgate 1, Oslo NØ. (1938).
Bryhni, Inge, cand. real. Institutt for geologi, Blindern, Oslo. (1955).
*Bryn, Knut Ørn, statsgeolog. Norges geologiske undersøkelse, Eilert Sundts gate 32. Oslo. (1956).
Brøndbo, Knut, bergingeniør, Orkla gruber. Løkken Verk. (1955).
*Braastad, Johan, dr. Postboks 808, Oslo. (1913).
Bugge, Arne, dr. Utsiktsveien 21, Stabekk. (1914).
*Bugge, Carl, direktør. Håkon den gode's vei 12, Vindern, Oslo. (S).
*Bugge, Jens A. W., professor. Geologisk Institutt, N.T.H., Trondheim. (1940).
*Bulman, O. M. B., professor. Sedgwick Museum, Cambridge, England. (1951).
Busé, M. L., Miss. Av. de las Americas 6-19, Zona 14, Guatemala City, Guatemala. (1960).
Bølviken, Bjørn, ingeniør. Statens Råstofflaboratorium, Postboks 1078, Trondheim. (1958).
*Callisen, Karen, dr. Mineralogisk og Geologisk Museum, Østervoldgade 7, København K, Danmark. (1917).
*Carstens, Harald, dr. Saxenborg, Trondheim. (1949).
Caster, Kenneth, professor. Dept. of Geology and Geography, University of Cincinnati, Cincinnati, Ohio, U.S.A. (1953).
*Chapman, Charleton, A., professor. University of Illinois, Urbana, Ill., U.S.A. (1949).
*Chen Kwan Yuan, B. Sc. Geological Institute, National Peking University, Peking, Kina. (1951).
Christensen, A. H., sivilingeniør, Lauritz Sands vei 14 C, Eiksmarka, Røa, Oslo. (1961).
Christie, Olav, cand. real. Geologisk museum, Oslo NØ. (1959).
Collini, Bengt, fil.lic., laborator. Kungl. Lantbrukshögskolan, Institutionen för Marklära, Uppsala 7, Sverige. (1949).
*Crowder, Dwight, dr. U.S. Geological Survey, 345 Middlefield Road, Menlo Park, Cal., U.S.A. (1958).
*Dahl, Eilif, dosent. Botanisk institutt, Norges Landbrukshøgskole, Vollebekk. (1946).
Dahl, Erling, ingeniør. Skogveien 1, Notodden. (1945).
Dahlström, Elis, dr. c/o I.I. G. Casilla 10465, Santiago, Chile. (1945).
*Danielsen, Anders, konservator. Botanisk museum, Bergen. (1949).
*Davidson, Charles F., professor, Department of Geology, St. Andrew University, Fife, Skotland. (1960).
*Devore, George W., dr. Dept. of Geology, Florida State University, Tallahassee, Florida, U.S.A. (1948).
*Dietrich, R. V., professor. Dept. of Geological Sciences, Virginia Polytechnic Institute, Blacksburg, Virginia, U.S.A. (1959).
Dietrichson, Brynjulf, bergingeniør. Hvitsten. (1917).
Divljan, Stevan, dr. Federal Geol. Institute ul. Rovinjskaja 12, Beograd, Jugoslavia. (1959).

Donnay, J. D. H., professor. Johns Hopkins University, Baltimore 18, Maryland, U.S.A. (1937).

*Dons, Johannes A., førstekonservator. Geologisk museum, Oslo NØ. (1941).

Dybdahl, Ivar, bergingeniør. A/S Titania, Hauge i Dalane. (1958).

*Eckermann, Harry von, professor. Edeby pr. Ripsa, Sverige. (1937).

*Edelman, Nils, dr. Bolidens Gruv AB., Boliden, Sverige. (1956).

Egeberg, F. P., bergingeniør. Munkedamsveien 91, Oslo. (1939).

Eggen, Johan, inspektør, Mo i Rana. (1961).

Einevoll, Ola, landbrukskandidat. Norges Landbrukshøgskole, Vollebekk. (1957).

Elder, T. G., geolog. Geologisk Museum, Sarsgt. 1, Oslo. (1960).

*Elders, Wilfred Allan, B. Sc. Dept. of Geology, University of Chicago, Chicago 37, Illinois, U.S.A. (1958).

Elliott, Robert B., dr. Department of Geology, The University, Nottingham, England (1954).

Englund, Jens O., stud. real., Institutt for geologi, Blindern, Oslo. (1961).

Enkovaara, Antti, magister. Oulainen, Finnland. (1947).

Eriksen, Edv. Andreas, bergingeniør. Munkedamsvn. 81b, Oslo. (1945).

Eriksson, Tryggve, chefsgeolog, c/o A. Johnson & Co., Stureplan 3, Stockholm 7, Sverige (1959).

*Eskola, Pentti, professor, Min.-geol. Inst., Universitetet, Helsingfors, Finnland. (1919).

Espeland, Margaret, B. Sc. Hus 2, Vei 3672, Nordstrandhøgda, Oslo. (1950).

*Falkenberg, Otto, dr., ingeniør. Priorveien 7, Smestad, Oslo. (1914).

Falkum, Torgeir, cand.mag., Nyquistveien 21, Norstrandhøgda (1961).

Faust, G. T., dr. U.S. Geological Survey, Washington 25 D.C., U.S.A. (1956).

Fernando, Leslie J. P., professor. Geological Survey & Mine Department, Colombo 2, Ceylon. (1960).

*Feyling-Hanssen, Rolf, statsgeolog. Institutt for geologi, Blindern, Oslo. (1950).

Filby, R. H., dr. Geologisk Museum, Sarsgt. 1, Oslo. (1960).

Flood, Boye, cand. real. Institutt for Geologi, Blindern, Oslo. (1960).

*Fuster, José M., professor. Museo National de Sciencias Naturales, Madrid, Spania. (1957).

*Fægri, Knut, professor. Botanisk museum, Bergen. (1935).

Færden, Johannes, statsgeolog. Norges geologiske undersøkelse, Trondheim. (1948).

*Føyn, Sven, rektor. Sørkedalsveien 93d, Oslo — Smestad. (1932).

Garthus, Torbjørn, lærer. Garthus, Valdres. (1943).

Gavelin, Sven, professor. Mineralogiska Institutet, Stockholms Universitet, Stockholm, Sverige. (1954).

Geis, Hans-Peter, dr. Rødsand gruber A/S, Rausand om Molde. (1958).

Gillberg, Gunnar, fil. dr. Hornsgatan 33b, Stockholm Sö, Sverige. (1952).

*Gjelsvik, Tore, direktør, dr. Norsk Polarinstitutt, Observatoriegt. 1, Oslo. (1945).

*Gjems, Odd, amanuensis. Norges Landbrukshøgskule, Vollebæk. (1958).

Gjessing, Just, professor. Geografisk institutt, Blindern, Oslo. (1953).

Gjessing, Leiv, læge. Engelsrud, Asker. (1945).

Gleditsch, Chr., overlærer. Kløftaveien 15B, Gjettum. (1941).

*Graff-Petersen, Poul, mag.scient. Mineralogisk Museum. Østervoldgade 7. København K., Danmark. (1957).

*Green, John C. dr. Department of Geology, University of Minnesota, Duluth 11, Minn., U.S.A. (1954).

Grenness, Johannes, ingeniør. Statens råstofflaboratorium, Østmarka, Trondheim. (1945).

*Grip, Erland, fil. lic. Boliden, Sverige. (1938).

Grønhaug, Arne, W., cand. real. Arupsgt. 10, Oslo. (1960).

*Grønlie, Arne, rektor. Mo komm. høgre almenskole, Mo i Rana. (1943).

Gustavson, Magne, cand. real., Norges geologiske undersøkelse, Trondheim. (1961).

Hafsten, Ulf, dosent. Botanisk Museum, Bergen. (1957).

*Hagemann, Fredrik, statsgeolog. Norges geologiske undersøkelse, Eilert Sundtsgate 32, Oslo. NV. (1951).

*Hagner, A. F., professor. Dept. of Geology, University of Illinois, Urbana. Ill., U.S.A. (1947).

Halfdansen, Olaf M., tegner. Grønseth & Co., Skøyen, Oslo. (1958).

*Hamar, Georg, vit.assistent. Institutt for Geologi, Blindern, Oslo. (1958).

Hansen, Kaj, dr. Gammeltoftsgade 16, København K., Danmark. (1947).

Hansen, Sigurd, dr. Danmarks geologiske undersøgelse, Charlottenlund, Danmark. (1960).

*Harry, W. T., B. Sc. Department of Geology, The University, St. Andrews, Fife, Scotland. (1949).

*Hartley, Jack, B. Sc. Department of Geology, University, Leeds 2, England. (1949).

*Hawkes, Leonard, professor, D. Sc., F.R.S. 26 Moore Lane, Rickmansworth, Herts., England. (1953).

*Heber, Gustav, advokat. Oscarsgt. 49, Oslo. (1937).

*Heber, Gustav, bygningsingeniør. Eikskollen 11, Røa, Oslo (1942).

*Heier, Knut S., dr. Department of Geology, Rice Institute, Houston 1. Texas. U.S.A. (1952).

*Heintz, Anatol, professor. Geologisk museum, Oslo NØ. (1926).

*Heintz, Natascha, konsulent. Norsk Polarinstitutt, Observatoriegt. 1, Oslo. (1951).

Heltzen, Anders M., bergingeniør. Fjellsprengningsutvalget, Løkkeveien 11, Oslo. (1941).

*Helverschou, Kaare, disponent. Villavegen 9, Frøen, Oslo. (1943).

*Henderson, Donald M., professor. Dept. of Geology, University of Illinois, Urbana, Il., U.S.A. (1949).

*Henningsmoen, Gunnar, førstekonservator. Geologisk museum, Oslo NØ. (1941).

*Hernes, Ivar, førstekonservator. Geologisk institutt. Bergen. (1947).

*Hesse, P. A. Jürgen, bergingenjör. Björkhagsvägen, Ormsta, Vallentuna, Sverige. (1936).

*Hessland, Ivar, professor. Stockholms Universitet, Stockholm, Sverige (1947).
*Hjelle, Audun, statsgeolog. Norges Geologiske Undersøkelse, Eilert Sundts gate 32. Oslo NV. (1958).
Hjelmquist, Sven, professor. Geologiska Inst., Lund, Sverige. (1949) .
Hofseth, Arne, professor. Norges Tekniske Høgskole, Trondheim. (1952).
Holmberg, Sigurd, amanuensis, Van Dürens väg 35A, Lund, Sverige. (1953).
Holmesland, P. B., overingeniør. Herøya pr. Porsgrunn. (1954).
Holmsen, Bjarne, ingeniør. Charlotte Andersens vei 28, Smestad, Oslo. (1941).
*Holmsen, Gunnar, dr. Skådalsveien 32. Vettakollen. (1908).
*Holmsen, Per, statsgeolog. Østhornveien 7. Nordberg, Oslo. (1940).
Holtedahl, Hans, dosent. Geologisk institutt, Bergen. (1948).
*Holtedahl, Olaf, professor. Geologisk institutt, Blindern, Oslo. (1908).
*Howe, Henry, V., professor. School of Geology, Louisiana State University, Baton Rouge, Louisiana, U.S.A. (1953).
Huseby, Fredrik, jernbanegeolog. Norges Statsbaner, Storgaten 33, Oslo.(1951).
Hvatum, O. Ø., geokjemiker, Statens Råstofflaboratorium, Trondheim. (1961).
Hysingjord, Jens, cand. real. Norges geologiske undersøkelse, Trondheim.(1959)
*Høeg, Ove Arbo, professor. Farmasøytisk Institutt, Blindern, Oslo. (1924).
Høegh-Omdal, S. K., byråsjef. Utgarveien 12, Jar pr. Oslo. (1943).
Ihle, Ole-K., lektor. St. Svithun skole, Stavanger. (1956).
Ingvaldsen, Karl, dir. Sverresgt. 4, Trondheim. (1956).
*Isachsen, Fridtjov, professor. Geografisk institutt, Blindern, Oslo. (1929).
Iversen, Anker, preparant. Geologisk institutt, N.T.H., Trondheim. (1953).
*Jakhelln, Johanne Hødal, cand. real. Solbakkevej 35, Gentofte, Danmark. (1943).
*Jakhelln, Kirsten, fru. Irisvegen 18, Berg, Oslo. (1953).
Johannesen, Brynhild Steinvik, cand. mag. Bygdøy allé 8, Oslo. (1959).
Johansen, Erling, stipendiat. Kråkerøy, Fredrikstad. (1952).
Jørgensen, Per, cand. mag., Institutt for geologi, Blindern, Oslo. (1961).
Jørstad, Finn, geolog. Norges geotekniske institutt, Blindern. (1957).
*Jøsang, Ottar, mag. scient. Geologisk Institutt, N.T.H., Trondheim. (1955).
Karlsen, J. U., verksmester, A/S Grønseth & Co., Skøyen, Oslo. (1960).
*Kautsky, Gunnar, statsgeolog. Sveriges Geologiska Undersökning, Stockholm 50, Sverige. (1952).
*Kennedy, W. Q., D., F. R. S., professor. University of Leeds, Leeds 2, England. (1947).
*Kiil, Erling, bergingeniør. Brantenborgveien 2, Slemdal, Oslo. (1922).
Kildal, Ellen S., cand. mag., Geologisk institutt, Bergen. (1960).
*Klompé, Theo, professor. Dept. of Geology, Faculty of Science, Chulalongkorn University, Phya Thai Rd., Bangkok, Thailand. (1947).
Klüver, Emil S., ingeniør. Høvik terrasse 38, Høvik. (1926).
*Kolderup, Niels-Henr., professor, Geologisk institutt, Bergen. (1919).
*Kollerud, Marta, cand. mag. Trondheimsveien 10, Oslo. (1929).
Kraft Johanssen, J., direktør. A/S Sydvaranger, Kongens gt. 5, Oslo. (1955).
*Krauskopf, K. B., professor. Geol. Dept. Stanford University, Stanford, Cal. U.S.A. (1953).

*Kristoffersen, K., dr. A/S Norsk Jernverk, Mo i Rana. (1941).
*Kullerud, Gunnar, dr. Geophysical Laboratory, 2801 Upton Street, Washington 8, D.C., U.S.A. (1948).
*Kulling, Oskar, statsgeolog. Sveriges Geologiska Undersökning. Stockholm 50, Sverige. (9153).
Kvale, Anders, professor. Geologisk institutt, Bergen. (1936).
Kvalheim, Aslak, laboratoriesjef. Statens råstofflaboratorium, Østmarka Trondheim .(1944).
Landergren, Sture, dr. Sveriges Geologiska Undersökning, Stockholm 50, Sverige. (1943).
Landmark, Einar, direktør. Tidemandsgt. 14, Oslo. (1941).
Landmark, Kåre, direktør. Tromsø museum, Tromsø. (1943).
*Larssen, Kari Egede, statsgeolog. Nils Bays vei 27A, Ullevål Hageby, Oslo. (1955).
*Larsson, Tollef, cand. real. Engene Fabrikker, Sætre i Hurum. (1945).
Laursen, Dan., professor, Jackson Jr. College, Jackson, Mich., U.S.A. (1947).
*Lien, Håkon, geolog. 144 Leo Ave., Sault Ste. Marie, Ontario, Canada. (1955).
Lien, Josef, ingeniør, cand. real. Bjørnsonsgt. 16, Bergen. (1959).
*Liestøl, Gerd Brevig, vit.assistent. Institutt for Geologi, Blindern, Oslo. (1958).
Liestøl, Olav, glasiolog, Norsk polarinstitutt, Observatoriegt. 1, Oslo. (1947).
*Lindley, Henry, W., dr. Die techn. Hochschule, Wullner Str. 2, Aachen, Tyskland. (1928).
*Ljunggren, Pontus, docent. Naciones Unidas, Casilla 686, La Paz, Bolivia. (1952).
Logn, Ørnulf, lic. techn. Fosdalen Bergverks A/S, Malm. (1953).
Lund, Kjell, direktør. Kvernveien 7, Øvre Ullern — Oslo. (1938).
*Lund, Lars, mag. sc. Forsvarets forskningsinstitutt, Kjeller pr. Lillestrøm. (1943).
Lund, Worm, bergingeniør. A/S Sydvaranger, Kongensgt. 5, Oslo. (1941).
*Lundby, Sven Erik, arkitekt. Norges byggforskningsinstitutt, Blindern, Oslo (1937).
Løddesøl, Åsulv, direktør. Det norske myrselskap, Rosenkrantzgt. 8, Oslo. (1938).
Løkås, Bjørge, lektor. Harry Borthensvei 8, Østmarkneset, Trondheim. (1960).
Løvaas, Gunnar, bergingeniør. Skorovas Gruber, Skorovatn p. å. (1949).
*Låg, J., professor, Norges landbrukshøgskole, Vollebekk. (1943).
Mackenzie Scott, Wm. Dept. of Geology, Manchester 13, England. (1955).
Magnusson, N. H., överdirektör. Vallhallavägen 46, Stockholm, Sverige. (1937).
*Major, Harald, geolog. Norsk polarinstitutt, Observatoriegt. 1, Oslo. (1944).
Marklund, Nils, fil. lic. Geologiska Institutet, Stockholms Universitet, Stockholm VA. (1951).
*Marmo, Vladi, Direktør. Geologiska Forskningsanstalten, Helsingfors. Finnland. (1949).
*Marthinussen, Marius, lektor. Ski komm. h. almenskole, Ski. (1931).
*Mason, Brian, dr. The American Mus. of Nat. History, Central Park West at 79th Street. New York 24, N. Y., U.S.A. (1953).

McCulloh, Thane, H., M. Sc. Division of Physical Sciences, University of California, Riverside, Cal. U.S.A. (1951).
Meier, Otto, dr. chefsgeolog. Svenska Diamantborrnings A/B, Hamngatan 29, Sundbyberg, Sverige. (1959).
Mellis, Otto, professor. Min. inst. Stockholms Universitet, Stockholm, Sverige. (1951).
[1]Michot, Paul, professor, Inst. de Géologie, Université de Liège, Belgia. (1951).
Mortensen, Odd, direktør. Postboks 744, Bergen. (1938).
Mortenson, Magne, professor. Norges tekniske høgskole, Trondheim. (1948).
Morton, Roger D., Ph. D. Dept. of Geology, The University, Nottingham, England. (1959).
Munck, Sole, museumsinspektør. Mineralogisk og Geologisk Museum, Østervoldgade 7, København K, Danmark (1948).
*Murray, Grover, E., professor. School of Geology, Louisiana State University. Baton Rouge, Louisiana, U.S.A. (1953).
*Naterstad, Johan, cand. real. Geologisk Institutt, Bergen. (1959).
Näykki, Ossi, fil. kand. Pihlajatie 43A, Helsingfors, Finnland. (1956).
*Neumann, H., førstekonservator. Geologisk museum, Oslo NØ. (1940).
Nikolaisen, Frank, stud. real., Geologisk museum, Sars gate 1, Oslo NØ. (1961).
*Nilssen, Borghild, cand. mag. Geologisk museum, Oslo NØ. (1955).
*Noe-Nygaard, Arne, professor. Østervoldgate 7, København K., Danmark. (1934).
Norin, Erik, professor. Mineralogisk-geologisk Institution, Uppsala, Sverige. (1947).
*Nosov, Edmund, dr. Box 282, Kentucky Geological Survey, Hopkinsville. Kentucky, U.S.A. (1953).
*Oftedahl, Christoffer, professor. Geologisk institutt, N.T.H., Trondheim. (1941).
*Oftedahl, Ivar, professor. Institutt for Geologi, Blindern, Oslo. (1918).
Ormestad, Solveig Bonde, cand. jur. Bernhard Herresvei 19, Smestad, Oslo. (1959).
*Orvin, A. K., direktør. Håkon den godes vei 21, Vindern, Oslo. (1913).
*Padget, Erna, kjemiker. Stallvägen 26, Roslags-Näsby, Sverige. (1955).
*Padget, Peter, dr. Sveriges geologiska undersökning, Stockholm 50, Sverige. (1954).
Page, Lincoln R., dr., Melvin Village. New Hampshire, U.S.A. (1960).
Paulsen, Nils, bergingeniør. Berg, Maridalen, Oslo. (1947).
Pehkonen, Eero, geolog. Outo-kompu Oy, Outokompu, Finnland. (1955).
*Pettersson, Adam, direktør. Lysaker ved Oslo. (1918).
*Poulsen, Arthur O., statsgeolog. Niels Henrik Abels vei 31, Blindern, Oslo. (1941).
Prost, André, dr. Laboratoire de Géologie du S. P. C. N., 12 Rue Cuvier, Paris 5e, Frankrike. (1959).
*Quensel, Percy, professor. Stockholms Universitet, Stockholm, Sverige. (1916).
*Ramberg, Hans, professor. Dept. of Geology, Univ. of Chicago, Chicago 37, Ill., U.S.A. (1941).

Ramsay, Donald M., Ph. D. Department of Geology, Queen's College, Dundee, Skotland. (1960).
Ramsli, Gunnar, Universitetslektor. Geografisk Institutt, Blindern. Oslo. (1948)
*Randall, B.A.O. Department of Geology, Kings College, Newcastle, England. (1958).
Rankama, Kalervo, professor, Geologisk Institut, Universitetet, Helsingfors, Finnland. (1947).
Rao, S. V. L., ass. lecturer in mineralogy. Indian Institute of Technology, Kharagpur, India. (1956).
Regnéll, Gerhard, professor. Geologisk-paleontologiska Inst., Lund, Sverige. (1947).
*Reitan, Paul H., Dept. of Geology, Stanford University, Stanford, Cal., U.S.A. (1956).
*Rosendahl, Halvor, konservator. Lindhaugen, Gyssestad, Sandvika. (1918).
*Rosenlund, A. L., jernbanegeolog. Drammensveien 103, Oslo. (1912).
Rosenqvist, Annemor, førstekonservator. Oldsaksamlingen, Historisk museum, Oslo. (1961).
Rosenqvist, Ivan Th., professor. Institutt for geologi, Blindern, Oslo. (1941).
Rowell, A. J., dr. Dept. of Geology, The University, Nottingham, England. (1959).
*Rønning, O., ingeniør. Konvallvn. 52, Oslo. (1938).
Sahlin, Anders, ingenjør. L. M. E., Stockholm 32, Sverige. (1953).
Samuelsen, Andreas, lektor. Lerkeveien 30, Bergen. (1929).
*Sandvik, Per Munthe-Kaas, bergingeniør. Orkla gruber. Løkken Verk. (1958).
*Savolahti, Antti, fil. kand. Jääkärinkatu 10A. Helsingfors, Finnland. (1956).
*Scheuman, K. H., professor. Wigbertstrasse 9, Bad Hersfeld, Tyskland. (1928).
Sellevoll, M. A., dosent. Jordskjelvstasjonen, Villaveien 9, Bergen. (1954).
Selmer-Olsen, Rolf S., førsteamanuensis, Geologisk Institutt, N.T.H., Trondheim. (1948).
Semb, Gunnar, forsøksleder. Norges landbrukshøgskole, Vollebekk. (1941).
*Siggerud, Thor, geolog. Norsk Polarinstitutt, Observatoriegt. 1, Oslo. (1951).
*Simpson, E. S. W., professor. Dept. of Geology, University of Cape Town, Rondeborch, South Africa. (1960).
Singsaas, Per, geofysiker, Geofysisk Malmleting, Trondheim. (1951).
Skjerlie, Finn, amanuensis, Geologisk Institutt, Bergen. (1951).
*Skjeseth, Steinar, statsgeolog. Norges geologiske undersøkelse, Eilert Sundts gate 32, Oslo NV. (1949).
Skordal, Asbjørn, geolog. «Vestråt». Elvegaten 18, Trondheim. (1946).
Skålvoll, Harald, cand. real. Norges geologiske undersøkelse, Trondheim. (1956)
Sletthaug, Mikal, bergingeniør, Seterveien 14, Norstrandshøgda, Oslo. (1960).
*Smithson, S. B., geolog. Geologisk museum, Oslo NØ. (1960).
*Sobral, José M., dr. Avenida de los incas 3020, Buenos Aires, Argentina. (1931).
Solli, Roar, ingeniør, Norges geologiske undersøkelse, Trondheim. (1961).
Soot-Ryen, T., direktør. p.t. Zoologisk museum, Oslo NØ. (1946).
*Spjeldnæs, Nils, førsteamanuensis, Institutt for geologi, Blindern, Oslo. (1946).
*Steinsvoll, Søren, overgartner. Botanisk hage, Trondheimsvn. 23, Oslo NØ. (1942).

Storebakken, H. C., lektor. Drammens Realskole, Drammen. (1940).
Strand, Geir, bergingeniør. Fosdalen Bergverk A/S, Malm. (1958).
Strand, P. O., sivilingeniør. Chr. Portland Sementfabrik, Slemmestad. (1959).
*Strand, Trygve, professor. Institutt for Geologi, Blindern, Oslo. (1927).
*Strøm, Kaare, professor. Limnologisk institutt, Blindern, Oslo. (1932).
Sturt, Brian A., Ph. D. Geology Department, Bedford College, Regent's Park, London N.W. 1. (1960).
*Størmer, Leif, professor. Institutt for Geologi, Blindern, Oslo. (1925).
Sund, Tore, dosent. Norges handelshøyskole, Bergen. (1943).
*Sverdrup, Thor Lorck, statsgeolog. Norges geologiske undersøkelse, Trondheim, (1955).
*Sverre, Erik, jr., disponent. Drammensvn. 44b, Oslo. (1946).
Svinndal, Sverre, geolog. Geofysisk Malmleting, Trondheim. (1953).
*Sweet, Walter, dr. Department of Geology, The Ohio State University, Columbus 10, Ohio, U.S.A. (1957).
Sæbø, Per Chr., cand. mag. Institutt for geologi, Blindern, Oslo. (1955).
Sæther, Egil, dr. Harestua. (1943).
*Sørbye, Ruth C., cand. real. Sollien 72D, Bergen. (1941).
*Sørensen, Henning, mag. scient. Mineralogisk Museum, Østervoldgade 7, København K, Danmark. (1952).
Tan, Tek Hong, geolog, Geofysisk Malmleting, Trondheim. (1961).
*Thorkildsen, Chr. Dick, cand. real. Norges geologiske undersøkelse, Trondheim. (1956).
Thorslund, Per, professor. Paleontologiska Institutionen, Uppsala 8, Sverige. (1940).
Tjølsen, Kjell, lektor, Eiganes skole, Stavanger. (1959).
Torske, Tore, cand. real. Geologisk Institutt, Universitetet i Bergen. (1959).
*Trøften, Einar, bergingeniør. Persauneveien 17, Trondheim. (1921).
*Trøften, Per-Fredrik, bergingeniør. Geofysisk Malmleting, Trondheim. (1955).
*Undås, Isak, lektor. Fløenbakken 8b, Bergen. (1931).
*Uytenbogaardt, W., professor. Geologisch Institut, Vrije Universiteit, De Lairessestraat 142, Amsterdam, Nederland. (1960).
Viluksela, Erkki, fil.kand. Outokumpu Oy, Outokumpu, Finnland. (1956).
Vokes, Frank Marcus, dr. Geologisk Museum, Sarsgt. 1, Oslo NØ. (1956).
*Wager, L. R., professor, D. Sc., F.R.S. The University Museum, Oxford, England. (1949).
[1]Wegmann, C. E., professor, Fgb. de l'hôpital 43, Neuchâtel, Schweiz. (1925).
Whitaker, J. H. Macdonald, Department of Geology, Leicester University, Leicester, England, (1961).
*Wickmann, Frans E., professor, Riksmuseet, Stockholm 50, Sverige. (1945).
Wilhelmsen, John, tekn. ass., Norges geologiske undersøkelse. Trondheim. (1956).
*Winsnes, Thore, geolog, Abbedikollen 34, Ullern. (1947).
Wolff, Fredrik Chr., Statsgeolog. Norges geologiske undersøkelse. Trondheim. (1958).
*Wright, Anthony D., dr. Department of Geology, Queen's University, Belfast, Nord-Irland. (1960).

Wærn, Bertil, fil. lic., bibliotekar, S:t Johannesg. 22, Uppsala, Sverige. (1947).
*Zachariasen, William, professor. Ryerson Physical Laboratory, University, Chicago 37, Ill., U.S.A. (1929).
*Zachrisson, Ebbe, fil. mag. Radarvägen 14, 2 tr., Viggbyholm, Sverige. (1958).
Ødegaard, Magne, bergingeniør, Statens Råstofflaboratorium, Trondheim. (1961).
Ödman, Olof H. R., professor. Inst. för mineralogi och geologi. Kungl. Tekn. Högskolan, Stockholm 70, Sverige. (1953).
*Øien, Fritz, bergingeniør, Midtåsen 50, Ljan. (1945).
*Ørvig, Tor, dr. Paleozoologiska Avd., Riksmuseet, Stockholm 50, Sverige. (1941).
*Øverlie, Olav, direktør. Rødsand gruber, Rausand om Molde. (1949).
*Ånerud, Kåre, lektor. Mossevegen 220, Nordstrand, Oslo. (1931).
Aasgaard, Gunnar, bergmester. Breiens vei 30, Nordstrandhøgda, Oslo. (1921).

2	korresponderende medlemmer.
141	livsvarige medlemmer.
153	årsbetalende medlemmer.
294	medlemmer.